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SURFACE WAVES: SOURCE AND PATH PROPERTIES

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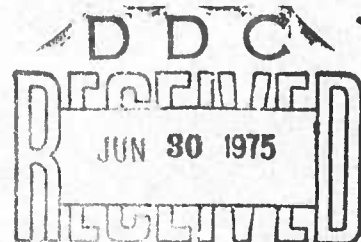
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20. ABSTRACT (Continue on reverse side if necessary and identify by block number) Several characteristics of oceanic surface waves can be altered by low rigidity sediments along the propagation path. Specifically spectral shape of both Love and Rayleigh waves as well as path phase velocity can be affected. The theoretical Love wave spectra is sensitive to both the nature of the low velocity zone and the sediment structure. Thus, observed Love wave spectra do not constrain focal depth.		

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## INTRODUCTION

We have been investigating the properties of seismic surface waves with the ultimate goal of using these waves to determine the mechanism and depth of the source. This information will be useful to discriminate explosions from earthquakes. To date, the major thrust of the research has been to investigate the effect of the propagation path on the surface wave. If the path effects cannot be removed from the observed signal then little information concerning the source can be determined.

We have examined the surface waves from two mid-Atlantic ridge earthquakes. These earthquakes, 2 June 1965 and 19 June 1970, have previously been well characterized by their Rayleigh wave signals (Weidner and Aki, 1973, J. Geophy. Res. 78, 1818.). In the present study, we mainly investigate the Love wave properties. The N-S and E-W long period records from a total of 35 WWSSN stations were digitized and the Love waves were separated from the horizontal Rayleigh wave by rotating the coordinate system to radial and transverse. These surface waves could then be processed in a computer. Our major conclusions are based on three types of information: 1) short period surface waves, 2) long period surface waves, 3) general properties.



## SHORT PERIOD SURFACE WAVES

Depth resolution for depths shallower than 10 km is dependent on surface waves analyses for periods shorter than about 20 sec. It is therefore important to understand how the propagation path can affect short period surface waves. We have made the following observations concerning short period surface waves:

1. Love waves are dispersive for periods shorter than about 20 sec. This dispersion coincides with the short period Rayleigh wave dispersion. However, the Rayleigh wave dispersion can result from the presence of the water layer as the Love wave propagation should be insensitive to this. We carefully analyzed the wave train to be sure that the transverse component had not been contaminated with the Rayleigh wave signal. The Love wave signal was correlated with and orthogonalized with respect to the horizontal Rayleigh wave signal. The dispersed wave train persisted and we conclude that it is a property of the Love wave..
2. The spectral amplitude shape of Love waves (which is largely controlled by the short period amplitudes) is highly variable from station to station. While it is difficult to predict the excitation amplitudes due to the dependence on details of the source medium model, we can conclude that there should be no azimuthal dependence of spectral shape. Thus, the observed variations must be due to path effects.

These Love wave properties were synthesized with various short period Rayleigh wave properties to derive conclusions about the

propagation path. The sediments along the path appear to be responsible for most of the short period observations. While these interpretations pertain to an oceanic path, they would be equally valid for any surface wave path which contained significant amounts of unconsolidated sediments.

Attached as Appendix A is a manuscript specifically addressing the effect of these sediments. The surface wave (Love or Rayleigh) with a period equal to  $4H/\beta$  (where  $H$  is sediment thickness and  $\beta$  is shear velocity) is strongly perturbed by the sediment layer. The surface wave becomes quite dispersed in the neighborhood of this period with a new higher mode branch being introduced at shorter periods. The eigenfunctions (of stress and displacement) became significantly distorted with a large percentage of the energy trapped in the sediment layer. Thus, the short period surface wave can be lost in transit due either to viscous attenuation in the sediment layer or to scattering from an impedance barrier. The models and observations which lead to these conclusions are discussed in detail in the Appendix.

#### LONG PERIOD SURFACE WAVES

While analyses of short period surface waves help resolve the depths of shallow sources, the long period surface wave can aid in determining whether or not the source is shallow. Furthermore, the

longer periods are less eroded by surficial path properties (like the sediments). Thus, while the resolution is reduced, the chance of seeing any source affect is greatly improved by studying the long period surface wave. In a previous study Weidner and Aki (J. Geophys. Res. 78, 1818, 1973) concluded that Rayleigh wave phase could be used to determine focal depths even in the absence of short period waves. In order to deduce the phase of the source, the phase delay of the path must be accurately known. The path delay can be calculated if the velocity and density structure of the medium is known. A significant piece of information which can be used to deduce path properties is the Love wave phase velocity. In many instances Love wave phase velocity may be easier to determine than Rayleigh wave phase velocity since both require a knowledge of the source phase and the Love wave source phase is much more independent of focal depth than the Rayleigh wave source phase. Thus, a potential approach to determining focal depth would be to first determine the Love wave focal phase, then the Love wave phase velocity. Next one would calculate the medium structure and Rayleigh wave phase velocity. Finally the path phase delay could be removed from the observed Rayleigh wave phase to determine the focal phase of the Rayleigh wave. This information could then be used to determine focal depth. The propagation of errors along with assumptions regarding isotropy are probably too large for unambiguous results. Nevertheless, restrictions



may be possible--we are currently pursuing this type of approach to a limited extent. We have calculated the Love wave phase velocity for the paths studied. We find that identical isotropic structural models describe both the Love wave and Rayleigh wave dispersion for many paths. We must now investigate the paths that do not presently appear to have a structure which is compatible with both data. Part of the problem may just be sensitivity to the properties at different depths for the different wave types. Further insight may be gained by studying the resolution kernels for the individual waves as well as a combined data set.

#### GENERAL PROPERTIES

Since the Love wave eigenfunctions for the medium studied vary only small amounts with period and depth over a broad depth range, the Love waves will not by themselves be useful in determining focal depth. Thus, Love waves may at best be used for a standard with respect to which the Rayleigh waves can be compared. We have already discussed the potential use of Love waves to aid in finding the Rayleigh wave source phase. Another aspect would be to use Love wave amplitudes to determine the Rayleigh wave source amplitude. There are two complementary methods of viewing the Rayleigh wave amplitudes which are

useful in defining focal depth. The first is viewing the amplitude as a function of frequency at a particular station. Source depth will be manifest as nodes or large amplitudes for various frequencies. The major problem is to be sure that the path attenuation is not dominating the amplitude spectrum. The second is to view the amplitudes as a function of azimuth. Here one must be concerned that lateral variations in the path properties do not dominate the source effects. In both cases, it is possible to use the Love wave amplitudes to reflect path properties. Our analyses suggest that the Love waves may be plagued with medium properties which differently affect the Rayleigh waves. The most serious pertain to the first type of depth analysis. We have calculated the excitation of Love waves with frequency. We find that the spectral shape of the Love wave is dominated by the detailed model of the low velocity zone and sediment cover in the source area. These two factors are difficult to know to the accuracy required. Thus, simply taking the ratio of observed Love to Rayleigh amplitude as a function of frequency may bear more on the source structure (through the Love wave) than on the focal depth (through the Rayleigh wave). Furthermore, sediments on the path will have somewhat different effects on the two wave types. Comparing the amplitudes of Love and Rayleigh waves as a function of both azimuth and frequency may be more useful. The excited Love wave amplitude depends on the source medium structure but the frequency dependence of the amplitude should be the same for all azimuths. Thus, by observing the lateral variations of the Love wave amplitudes, we may be able to separate the effects of path lateral variations from the radiation pattern of Rayleigh waves.

## Appendix A

### The Effect of Oceanic Sediments on Surface Wave Propagation

#### Abstract.

Several characteristics of oceanic surface waves can be altered by the presence of low rigidity sediments along the propagation path. Love and Rayleigh waves from mid-Atlantic ridge earthquakes bear many effects of oceanic sediments. The general absence of these surface waves for periods shorter than about 15 seconds can be attributed to either attenuation or scattering in thin sediments. Thin sediments also disperse short period Love waves. Sediments whose thickness exceeds about two kilometers are responsible for removing surface wave energy with periods up to 40 seconds. These sediments also alter the particle motion of Rayleigh waves and are responsible for a complicated dispersion relation. These thick sediments substantially reduce the surface wave phase velocity at periods in excess of 100 seconds.

## The Effect of Oceanic Sediments on Surface Wave Propagation

### Introduction.

Several features of observed surface waves can be used to deduce source and path properties. Tsai and Aki (1970) deduced focal depth from the amplitude spectral shape of Love and Rayleigh waves. Mendiguren (1971) was able to determine focal depth from the radiation pattern of Love and Rayleigh waves. Weidner and Aki (1973) found that Rayleigh wave phase could be used to constrain focal depth. Oceanic upper mantle structures have been deduced from single station phase velocity determinations by Weidner (1974), Leeds et al., (1974) and Forsyth (1973).

These types of studies assume that the observed surface wave character has not been significantly altered by indeterminate path properties which are unrelated to the investigated property. The physical properties of oceanic sediments are largely unknown yet they are ubiquitous in ocean environments. In this paper, the nature and magnitude of the effects of sediments on surface wave propagation is estimated. Theoretical calculations are compared with observations of both Love and Rayleigh waves in the Atlantic Ocean. We conclude that thin sediments control the short period amplitudes while thick sediments severely affect the dispersion to very long periods.

### Analytical Models

The presence of a low-rigidity sediment layer in an oceanic, velocity-density structure will strongly control many aspects of the surface wave character. Several features inherent to surface waves for a single low-rigidity layer over a half-space are directly related to similar features of a more complicated layered half-space with a superficial water layer. Several authors have analytically studied a variety of simple models appropriate to the discussions (Tazime, 1958, 1959, 1962; Ohta, 1962; and Mooney and Bolt, 1966). In the first of these papers Tazime (1958) investigated the nature of the  $M_1$  and  $M_2$  branches of the Rayleigh wave equation for a plate as the Poisson's ratio approached 0.5 (liquid case). He found that the fundamental mode of each branch did not uniformly approach the fundamental modes for a liquid plate. Instead, the fundamental modes of the liquid plate must be constructed from a superposition of the fundamental and higher modes of the solid. These branches of the wave equation for plates can be related to Rayleigh waves for a layer on a half-space (Ewing, Jandetzky and Press, 1957). In a similar fashion, as the shear modulus of a surface layer approaches zero, the fundamental Rayleigh wave does not approach that of a liquid layer overlying a half space. Again, the fundamental and higher modes must be superposed (Tazime, 1962). Figure 1 (from Tazime, 1962) illustrates this point. The solid curves are group velocities for Poisson's ratio,  $\sigma$ , in the layer of 0.48. The chain curves are for  $\sigma = 0.5$ .

Another interesting feature of the Rayleigh wave for a low rigidity layer over a half-space concerns the particle motion. Ohta (1962) points out that the



surface particle motion for the fundamental mode changes from retrograde to prograde and then to retrograde as period decreases. The frequency extent of the prograde region depends on the layer thickness and shear velocity. In addition, the vertical displacement for some frequencies has a zero crossing with depth. These features are generally associated with higher modes. However, the fundamental mode phase velocity is always lower than that of the higher mode in these cases. The higher modes may have retrograde surface motion and no zero crossings in the vertical displacement.

Mooney and Bolt (1966) suggest that the surface wave character depends mainly on the parameter  $\beta T/H$ , where  $\beta$  is the shear velocity of the low-rigidity layer,  $H$  is its thickness and  $T$  is period. Tazime (1962) finds large effects of the sediments on the surface wave dispersion and eigenfunctions for  $T\beta/H = 4/(2m+1)$  and  $T\alpha/H = 4/(2m+1)$ , where  $\alpha$  is the compressional velocity. These periods correspond to internal reflections in the low rigidity layer of S and P waves (Sykes and Oliver, 1964a). In Figure 1 the minimum for the group velocity of the fundamental mode corresponds to  $T\beta/H \approx 4$  since  $\frac{\alpha}{\beta} \approx 5$ . In addition, this period of minimum group velocity is longer than the cut-off period of the first higher mode. The phase velocities corresponding to the group velocities of Figure 1 are presented in Figure 2. These are also from Tazime (1962) and the different  $m$  correspond to the dispersion of Love waves when  $\mu_2/\mu_1$  become infinitely large. As we can see, the Rayleigh modes approach these curves.

### Effect of sediments on surface wave dispersion.

We have analyzed the Rayleigh waves from four mid-Atlantic ridge earthquakes and the Love waves from two of the events at several circum-Atlantic stations. The location of the events and other relevant information is given in Table 1 with a map in Figure 3. A map showing the sediment thickness in the North Atlantic is given in Figure 4 (Ewing et al., 1973).

The paths to TRN have a large portion of thick sediments. Weidner (1974) concluded that the Rayleigh wave phase velocity to stations around the Caribbean Sea was consistent with a sediment shear velocity of .5 km/sec. The sediments close to TRN for the path from event pair I are about 5 km. thick. These values suggest strong effects of the sediments for surface waves with 40 sec. periods. The theoretical Rayleigh wave dispersion for such a structure are illustrated in Figure 5. The phase and group velocity decrease sharply as the period decreases towards 40 sec. At shorter periods, the first shear mode exists and its group velocity initially increases sharply with decreasing period. We compare this phenomena with the observed group velocity obtained with the technique of Dziewonski et al., (1969) in Figure 7 where the observations for TRN and CAR for events 1 and 2 are illustrated. The contours represent equal energy levels. There is a very strong suggestion of an inversely dispersed branch arrival with the characteristic of the theoretical curves in Figure 5. These can be contrasted with more typical observations of BEC and BLA in Figure 6. The surface waves for other Cariggean stations (BHP, LPS) for both event pairs have a character similar to TRN and CAR for both Love and Rayleigh waves.

Another feature of the Rayleigh wave for the final ocean model to TRN is the presence of only prograde surface motion for all modes depicted in Figure 5 between the periods 28 to 40 sec. In Figure 8 are shown the observed difference between the **phase** of the vertical and horizontal components for three of the events at TRN. There is a consistent shift in the phase difference at about .025 Hz (40 sec). At longer periods the phase difference is close to  $\pi/2$ , indicating retrograde motion. Relating the theoretical value to the observed value is complicated by the presence of more than one mode. If a single mode is not isolated interference from the other mode can result. The observed relative phase shift at 40 sec. suggest that effects of the prograde motion are present, but the signal may be contaminated by higher modes. Other observations of this phase difference are often less clear than for TRN. The effect of lateral variations on these observations is difficult to describe. In addition, the surface wave particle motion should reflect the structure very close to the station. Thus, the type of phase shift that we suggest is present at TRN may not always be expected if the local sediment thickness is not uniform.

The group velocities for Love waves can be deduced from the energy contours in Figure 9 for some representative stations. Indeed, the Love waves are also dispersive at short periods. However, only the sedimentary layer can be responsible for this dispersion. These observations could result from contamination of the Love wave with horizontal Rayleigh waves by multi-path

Rayleigh wave propagation or simply an erroneous coordinate rotation of the seismogram. We have calculated the correlation coefficient of the Love and horizontal Rayleigh wave trains and parts of the wave trains. We then rotated the coordinate system to minimize the correlation and removed the dependent portion from the Love wave signal with a Gram-Schmidt technique. In all instances the dispersive nature of the Love wave remained. We thus conclude that the Love waves are dispersive and that the sediments are responsible. The coincidence in shape of the dispersion for Love and Rayleigh waves along with the absence of the Airy phase for recorded Rayleigh waves suggests that the sediments and not the water are controlling the short period Rayleigh wave dispersion.

#### Effect of sediments on surface wave amplitudes.

The amplitudes of Rayleigh waves from mid-ocean ridge earthquakes are generally low for periods shorter than 15 sec. This observation led Tsai (1969) to conclude that ocean ridge normal faulting events must be very deep (45-65 km) since shallow dip-slip events efficiently excite these short periods. The strike-slip events, on the other hand, could be shallow since the amplitude spectrum has a node at short periods. Weidner and Aki (1973) concluded that the four mid-Atlantic ridge earthquakes of Table 1 were shallow. The conclusion was primarily based on phase information.

The Rayleigh wave amplitudes for these events as a function of frequency for some stations are compared with the theoretical values in Figures 10 and 11. Here the displacement spectral density of the unfiltered record are displayed.

Corrections for instrument and geometrical spreading have been included. The high frequencies are not observed for the dip-slip events (events 1 and 3) even though they should have been generated. The amplitudes for the strike-slip events often do not agree at short periods, but small variations in the focal properties would correct this. Of the observations illustrated in Figures 10 and 11, BEC recorded the largest amplitude ratio of high frequency to low with ATL and BLA intermediate and BHP and CAR lowest.

The amplitudes of Love waves are compared with the theoretical values in Figure 12. The theoretical values are for the fundamental mode assuming that the medium can be described by the normal ocean basin of Weidner (1974). The frequency dependence of the theoretical amplitude is very sensitive to small changes in the medium structure. Even though the frequency dependence of the source amplitude is uncertain, it should be the same for all stations and events studied here. The lateral variations in the observed Love wave spectral shape must be attributed to path effects. The reliability of each observation can be made by comparing the Love waves from the two events at the same stations since the path should be the same. The observations in Figure 12 indicate a progressive erosion of high frequencies proceeding from BEC to ATL to BHP. The coincidence of eroded high frequency Love waves with similar features of the Rayleigh waves suggests that the spectral shape of both types of waves is a path effect.

Analyses of the Rayleigh waves for cross-Atlantic paths reveal that the short periods are lost in transit. Relevant maps for the paths analyzed are shown in Figures 13 and 14. For each of the four events, the Rayleigh waves



recorded at two stations along a great circle path were examined. The Rayleigh wave amplitudes are shown in Figure 15. It is very clear that the amplitudes at short periods were lost between the stations. We conclude that the low amplitudes for short-period Rayleigh waves from mid-ocean ridge earthquakes do not contradict shallow focal depths.

We suggest that sediments may be responsible for the loss of short-period surface waves. The mechanism can be either attenuation related to the low  $Q$  of sediments or reflection due to the change in the displacement and stress eigenfunctions as the sediment thickness and properties change.

We have investigated the effect of various thicknesses and shear velocities of a sedimentary layer on the surface wave. The crust and upper mantle is described by the normal ocean basin model (Weidner, 1974). As discussed earlier, the surface wave is affected for periods in the neighborhood of  $T_0$  where

$$T_0 = \frac{4H}{\beta}$$

$H$  being thickness of sediments with shear velocity,  $\beta$ . Sediment models with  $T_0 = 10, 15$  and  $30$  sec are examined. The models are described in Table 2. The group and phase velocities for these models are illustrated in Figures 16 and 17 for Rayleigh and Love waves. These curves were generated numerically and it was not always possible to follow a single mode, since neither polarization nor the number of zero crossings of the eigenfunctions can be used as diagnostics.

The depth dependence of the Rayleigh wave vertical displacement eigenfunctions is shown in Figure 18 for several periods. The eigenfunction for the 17 sec period is representative of the sediment free case. The displacement eigenfunction for Love waves is illustrated in Figure 19. The sediments have very significant effects. There are no Rayleigh waves in the period range 15.5 to 16.3 sec that have eigenfunctions resembling the sediment free case. Thus the surface waves in this period range will experience a large impedance mismatch when they travel from a sediment-free region with sediments described by this model. The breadth of the period range where the sediments alter the eigenfunction appears to be larger for Love waves than for Rayleigh waves.

Models I, III and IV give rise to similar variations in the eigenfunctions for periods close to the critical period. In general, the eigenfunctions are significantly altered relative to the sediment free eigenfunctions in the same period range where the group or phase velocities deviate from those of the simple model.

Assuming that the heterogeneous medium can be described by two infinite laterally homogeneous quarter spaces in contact, an impedance mismatch will result if the stress and displacement distributions are different for the two structures. The band width of the Rayleigh wave which is severely affected by the sediments does not exceed 1 sec for the models considered. From equation 1, the eigenfunctions for a given period will be affected by the sediments of thickness  $H_0 \pm \frac{\delta}{8}$ . The total thickness variation which will affect a

given period cannot exceed 10-100 meters. As can be seen in the isopach map in Figure 7, small variations in sediment thickness is accomplished over a very short distance for the deeper sediments and somewhat longer distances for shallow sediments. For the above assumptions to be valid, the width of the region affecting a given period Rayleigh wave must be large compared with the wavelength. Indeed, this assumption is not valid for most periods. For 15 sec waves, the wavelength is about 60 km, while the wavelength of 10 sec waves is only 20 km. An impedance mismatch should be more effective in eliminating the short periods than the longer periods. We conclude that it may be possible to scatter short-period waves and transmit long-period waves into higher modes as observed at TRN. Since the affected bandwidth for Love waves is longer, this mechanism will be more effective in removing Love waves than Rayleigh waves.

The absence of short-period surface waves may also be attributable to attenuation associated with low Q sediments. The effective quality factor of Rayleigh waves,  $Q_R$ , can be related to the quality factor for shear waves of a layer,  $Q_s$ , by the relation

$$Q_R^{-1} = Q_s^{-1} \left[ \frac{\beta}{c} \frac{\partial c}{\partial \beta} + \frac{4}{3} \left( \frac{\beta}{\alpha} \right)^2 \frac{\alpha}{c} \frac{\partial c}{\partial \alpha} \right] \quad (2)$$

where  $\beta$  is shear velocity,  $\alpha$  is compressional velocity and the partial derivatives are defined for a uniform layer (Anderson *et al.*, 1965). Then the amplitude at distance  $x$  is given as

$$A(x) = A_0 e^{-\frac{\pi}{TU} \frac{x}{Q_R}} \quad (3)$$

where T is period and U is group velocity. Similarly the quality factor for Love waves is given by

$$Q_L^{-1} = Q_S^{-1} \frac{\beta}{c} \frac{\partial c}{\partial \beta} \quad (4)$$

Using a computer program of Saito (Saito, 1967), we calculated the partial derivatives for the sediment models of Table 2. The ratio  $A/A_0$  for models I, II and IV are given in Figures 20 and 21 for various values of  $x/Q_S$ . We used the surface wave modes that most nearly approximate the modes of a sediment-free ocean for these calculations. When a lower mode exists, it has much more energy in the sediments and is attenuated faster. The calculations may not be precise at the period critically disturbed by the sediments since the eigenfunctions will vary considerably with small changes in properties.

Models I and IV have the same shear velocities and their corresponding  $T_0$ 's are on different sides of the Rayleigh wave Airy phase. The model with  $T_0$  on the short period side of the Airy phase is much more effective in attenuating the Rayleigh wave. In addition the period range affected is broader. This observation suggests that the short periods can be attenuated more effectively than the longer periods for a given sediment shear velocity. Next compare models I and II. Here the  $T_0$ 's are the same, only the thickness and shear velocity are changed. Thicker sediments are

much more effective in removing the short periods. The energy loss for Love waves appears to depend mainly on the sediment thickness.

There is strong evidence that suggests that the  $Q_s$  of sediments is of the order of 10-20. Kudo and Shima (1970) measured the attenuation of shear waves in soil in situ to depths of 40 meters. They found that the  $Q_s$  was fairly frequency-independent between 30 and 80 Hz with indications of decreasing  $Q_s$  with decreasing frequency. For different soils, the value of  $Q_s$  ranged from 5 to 20. Tullos and Reid (1969) measured the attenuation of P-waves in sediments in situ to depths of 1000 feet. The  $Q$  for P waves ranged between 70 and 200 for 50-400 Hz. The resulting  $Q_s$  is equal to  $\frac{4}{3} Q_p \left(\frac{\beta}{\alpha}\right)^2$  if the loss mechanism is due to shear only. Boore and Warrick (1972) reported a sediment  $Q_s$  from short-period Rayleigh waves in the San Francisco Bay of 10-25.

The ratio of shear velocity to sediment thickness appears to be a dominating variable. Reported sediment shear velocities range from .05 to 0.7 km/sec. Sykes and Oliver (1964b) modeled the sediments in the Argentine Basin with a shear velocity of .2 to .4 km/sec in the upper 1/2 km and .5 to .7 km/sec below this layer from Love and Rayleigh dispersion. Anderson and Latham (1969) concluded, from observations of the first shear mode, that the average shear velocity for 150-meter thick sediment layers in the Atlantic was .075 km/sec. Davies (1965), using Stoneley waves, obtained shear velocities increasing from .05 to .19 km/sec in the upper 16



meters of sediments in the Indian ocean. Hamilton et al., (1970) estimated shear velocities of .1 to .2 km/sec from Stonely waves. SH wave velocities obtained by Kudo and Shima (1970) for soil range between .1 and .4 km/sec. These values may be applicable to saturated sediments as Nur and Simmons (1969) have shown that the presence of water does not affect the shear velocities in rocks. These measurements are consistent with the values of shear velocities that we use in our models. Furthermore, the low Q for the sediments along with the curves in Figures 20 and 21 suggest that attenuation will be a significant factor in removing the short period surface waves.

The short-period surface wave amplitudes for four mid-Atlantic ridge earthquakes can be compared with the theoretical values in Figures 10-12. The short period Rayleigh waves recorded at BEC for the pair I events both exhibit a minimum in the amplitude at about .08 Hz with more energy at both longer and shorter periods. The 16 November 1965 recording at BEC is typical of most observations in that it does not indicate the increase in energy at the high frequencies. On the other hand, the 17 May 1964 event does not lack energy at these short periods. Surface waves from pair I events travel a great distance with very thin sediments and then the sediments thicken rather quickly (Figure 4). The 16 November 1965 path indicates a more gradual increase in sediment thickness over the entire path while the path from 17 May 1964 has a very sharp increase in thickness quite close to the event. As we have suggested, the path with the most gradual increase in sediment thickness (16 November 1965) has lost the high frequencies over the widest bandwidth.

The pair I event with the thin sediment layer lost energy over a narrower bandwidth. On the other hand, the waves from 17 May 1964, which experienced virtually no sediments thinner than 200 m, does not appear to have lost any high frequencies. The contours of the energy arrivals at short periods at BEU for the two events of pair II are given in Figure 22. The 16 November 1965 has a very well defined region where the group velocity is very steep. The 17 May 1964 record does not indicate this portion of the curve. Instead these short-periods are arriving at a much faster group velocity. We conclude that Rayleigh waves were able to transform from the fundamental mode to the first shear mode since the sediment thickness changed so rapidly along this path.

Three types of sediment regions can be defined with differing effectiveness of each mechanism in eliminating the high frequencies. The first region has a very flat sedimentary layer where the thickness varies quite slowly over several wavelengths. Reflection will not be very effective with the major loss of energy at periods in the neighborhood of  $T_0$ . If the sediments are thick, then the band width of the attenuated energy will be large extending from  $T_0$  to shorter periods. If  $T_0$  is less than the period of the Rayleigh wave Airy phase, the bandwidth will again be large. Rayleigh waves will be more severely affected than Love waves.

The second region is one where the sediments increase in thickness more rapidly over several wavelengths. In such a case, the eigenfunctions do change suddenly and the length of path with anomalous eigenfunctions is

at least a few wavelengths. These conditions give rise to significant loss of energy due to reflection. Attenuation will also be effective with the same characteristics described above. In this region short period Love waves will be more effectively removed than Rayleigh waves.

The third case is characterized by even larger changes in sediment thickness with distance. Now the higher modes are generated within a wavelength of the region where the fundamental mode eigenfunctions change in character. These higher modes exhibit eigenfunctions very similar to the normal fundamental mode and the energy can easily be transferred from one to the other. In this medium, attenuation will not be important as the previous cases since attenuation requires a reasonably long path with a constant  $T_0$ .

Some waves travelled great distances through thick sediments. Such stations include TRN, CAR, LPS, etc. In these cases, for all events, the short periods have been eroded very effectively. This is consistent with attenuation due to the thick sediments.

We can estimate the shear velocity of the sediments if we assume that the corner frequency corresponds to  $T_0$  and knowing the sediment thickness from the isopach map. The corner frequency for BEC does not significantly differ from most of the observations in eastern U. S. (ATL, SHA, BLA). Thus, it appears that most of the short periods are lost well out in the ocean. The sediment thickness, as indicated by the isopach map of Figure 4, does not exceed .5 km in this region. However, Ewing and Ewing (1959) have reported refraction data indicating thicknesses of as much as 1 km east of

Bermuda. For a  $1\frac{1}{2}$  km thickness, a corner frequency of 15 sec would correspond to a shear velocity of about .13 km/sec. As the sediments thicken to the west of Bermuda, their shear velocities and Q may also increase.

Short period fundamental mode Rayleigh waves have been observed in the Pacific. Mendiguren (1971) reported large amplitudes for 13 sec Rayleigh waves at GIE for a shock in the Nazca plate. Forsyth (personal communication) observed large amplitudes at the same station for 12 sec Rayleigh waves. These periods are shorter than those for most of the Atlantic observations. The event occurred Oct. 12, 1964 on the east-Pacific rise about 4000 km from GIE. Ewing et al. (1969) present an isopach map of the south Pacific. The path to GIE from these events appears to be reasonably devoid of sediments except very close to the station where they may accumulate to .2 km thickness.

## CONCLUSIONS

Oceanic sediments can have a significant effect on the surface wave character. Thick sediments such as in the Caribbean region can affect the Rayleigh wave phase velocity by as much as 0.06 km/sec for waves with a period of 100 sec (Weidner, 1974). Such thick sediments can also alter the sense of ground motion and introduce higher mode Love and Rayleigh waves of periods as long as 40 seconds. These thick sediments trap the surface wave energy and are capable of efficiently removing high frequencies by attenuation. In addition the high frequencies may be scattered as the sediment thickness varies.

Thin sediments appear to be responsible for the general absence of short period surface waves from mid-Atlantic ridge earthquakes. The loss may be either due to attenuation or an impedance barrier to high frequencies due to sediment thickness changes. These sediments disperse short period Love waves and are probably responsible for the short period Rayleigh wave dispersion.

The surface wave whose period is  $\frac{4H}{\beta}$  is most severely affected by the sediments. The effect is generally manifest by very small group velocities, most of the energy trapped in the sediment layer, considerably distorted depth dependence of displacement and stress, and the introduction of a higher mode of branch at shorter period. The character of the higher mode branch approaches that of the undisturbed fundamental mode as period decreases further. The band width where the surface wave is disturbed is generally larger for thicker sediments, larger (for Rayleigh waves) for periods



shorter than the Airy phase, and larger for Love waves than Rayleigh waves when the affected period is longer than the Airy phase.

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TABLE 1

Summary of Earthquake Locations and Other pertinent Data

Event Number	Date	Origin Time			Latitude	Longitude	Magnitude*
		h	m	s			
1	2 June '65	23	40	22.5	15.96°N	40.79°W	5.6
2	19 June '70	14	25	18.4	15.4°N	45.9°W	5.5
3	16 Nov. '65	15	24	40.8	31.03°N	41.49°W	6.0
4	17 May '64	19	26	16.4	35.29°N	36.07°W	5.6

Event Number	Seismic Moment $\pm$ (dyne cm)	Focal mechanism <sup>+</sup>			Reference for location and body wave 1st motion
		Strike	Dip	Slip	
1	$7.82 \times 10^{24}$	$15^\circ \pm$	$50^\circ \pm$	$-100^\circ \pm$	Sykes (1970)
2	$1.03 \times 10^{25}$	$8^\circ \pm$	$87^\circ \pm$	$160^\circ \pm$	US Coast and Geodetic Survey
3	$1.2 \times 10^{25}$	$0^\circ \pm$	$121^\circ \pm$	$54^\circ \pm$	Sykes (1967)
4	$1.94 \times 10^{25}$	$0^\circ \pm$	$76^\circ \pm$	$6^\circ \pm$	Sykes (1967)

\* Body wave magnitude, U. S. Coast and Geodetic Survey.

+ Convention of Tsai and Aki (1970).

+ From surface wave analysis of Weidner and Aki (1973).



Table 2 Sediment models

Number	Thickness (km)	Shear velocity (km/sec)	$T_o$ (sec)
I	0.1875	0.05	15
II	1.875	0.5	15
III	0.375	0.05	30
IV	0.125	0.05	10

### Figure Captions

- Figure 1 Group velocity of the Rayleigh wave modes for a layer over a half-space. The layer's Poisson ratio is 0.48, the half-space is 0.25. The densities are equal with a compressional velocity ratio of 4. The dashed line corresponds to a liquid over a half-space. (Tazime, 1962). The numbers refer to different modes.
- Figure 2 Phase velocities of Rayleigh wave for a layer over a half-space same model as Figure 4.1. The "m" curves correspond to the dispersion of Love waves when the shear modulus ratio of the layer and half-space are infinite. (Tazime, 1962).
- Figure 3 Locations of earthquakes and stations. The curves are representative great circle paths.
- Figure 4 Sediment thickness in the North Atlantic. Contours are in 100s of meters. (Ewing et. al., 1973).
- Figure 5 Rayleigh wave phase and group velocity for a normal ocean basin structure with 5 km of sediments whose shear velocity is 0.5 km/sec.
- Figure 6 Equal energy contours of the vertical Rayleigh wave component as a function of period and group velocity.
- Figure 7 Equal energy contours of the vertical Rayleigh wave component as a function of period and group velocity.
- Figure 8 Phase difference between vertical and horizontal component of the Rayleigh wave. The open circles are from the Fourier transform of the entire record; the crosses are from the time variable filtered record.
- Figure 9 Equal energy contours of Love waves as a function of period and group velocity.
- Figure 10 Rayleigh wave displacement spectral density. The observed values have been corrected for instrument and geometrical spreading. The theoretical values are for the depths and mechanisms of table 1.
- Figure 11 Rayleigh wave displacement spectral density. The observed values have been corrected for instrument and geometrical spreading. The theoretical values are for the depths and mechanisms of table 1.
- Figure 12 Love wave displacement spectral density. The observed values have been corrected for instrument and geometrical spreading. The theoretical values are for the depths and mechanisms of table 1.

- Figure 13 Maps for cross-Atlantic paths. Triangles indicate event locations and circles give the station locations.
- Figure 14 Cross-Atlantic paths. Triangles represent earthquakes and circles are stations.
- Figure 15 Rayleigh wave amplitudes for non-Atlantic earthquakes. Observed values have been corrected for instrument and geometrical spreading.
- Figure 16 Rayleigh wave phase and group velocities for different sediment models.
- Figure 17 Love wave phase and group velocities for different sediment models.
- Figure 18 Rayleigh wave vertical displacement eigenfunctions for sediment model 2 of table 2 at different periods. The values are normalized to unity at the surface.
- Figure 19 Love wave displacement eigenfunctions for sediment model 2 of table 2 and different periods and for a sediment free case. The values are normalized to unity at the sediment surface.
- Figure 20 Predicted ratio of observed to excited Rayleigh wave amplitude for different sediment models and values of  $x/Q$  when  $x$  is the propagation distance in kilometers and  $Q$  is the shear wave quality factor.
- Figure 21 Predicted ratio of observed to excited Love wave amplitude for different sediment models and values of  $x/Q$  where  $x$  is the propagation distance in kilometers and  $Q$  is the shear wave quality factor. The curves correspond to  $x/Q = 10$  and  $100$ .
- Figure 22 Equal energy contours of the BEC records viewed in the period vs. group velocity plane.

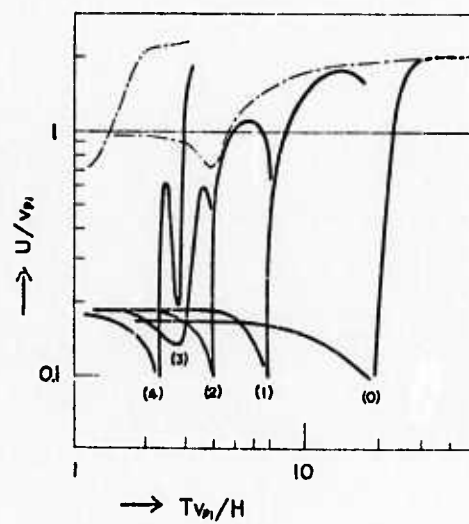


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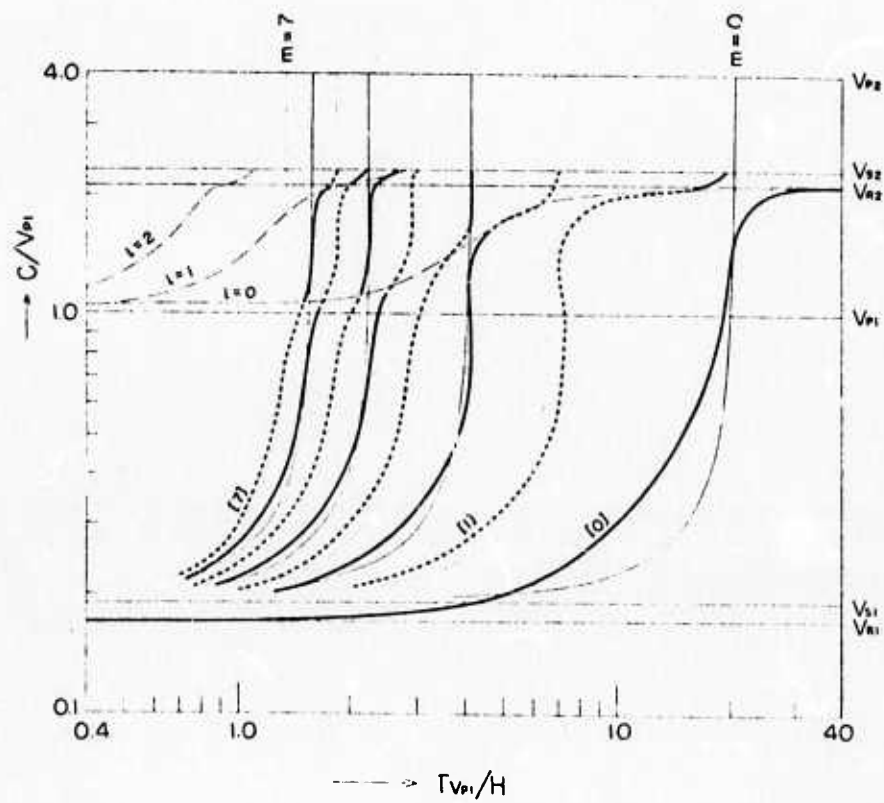


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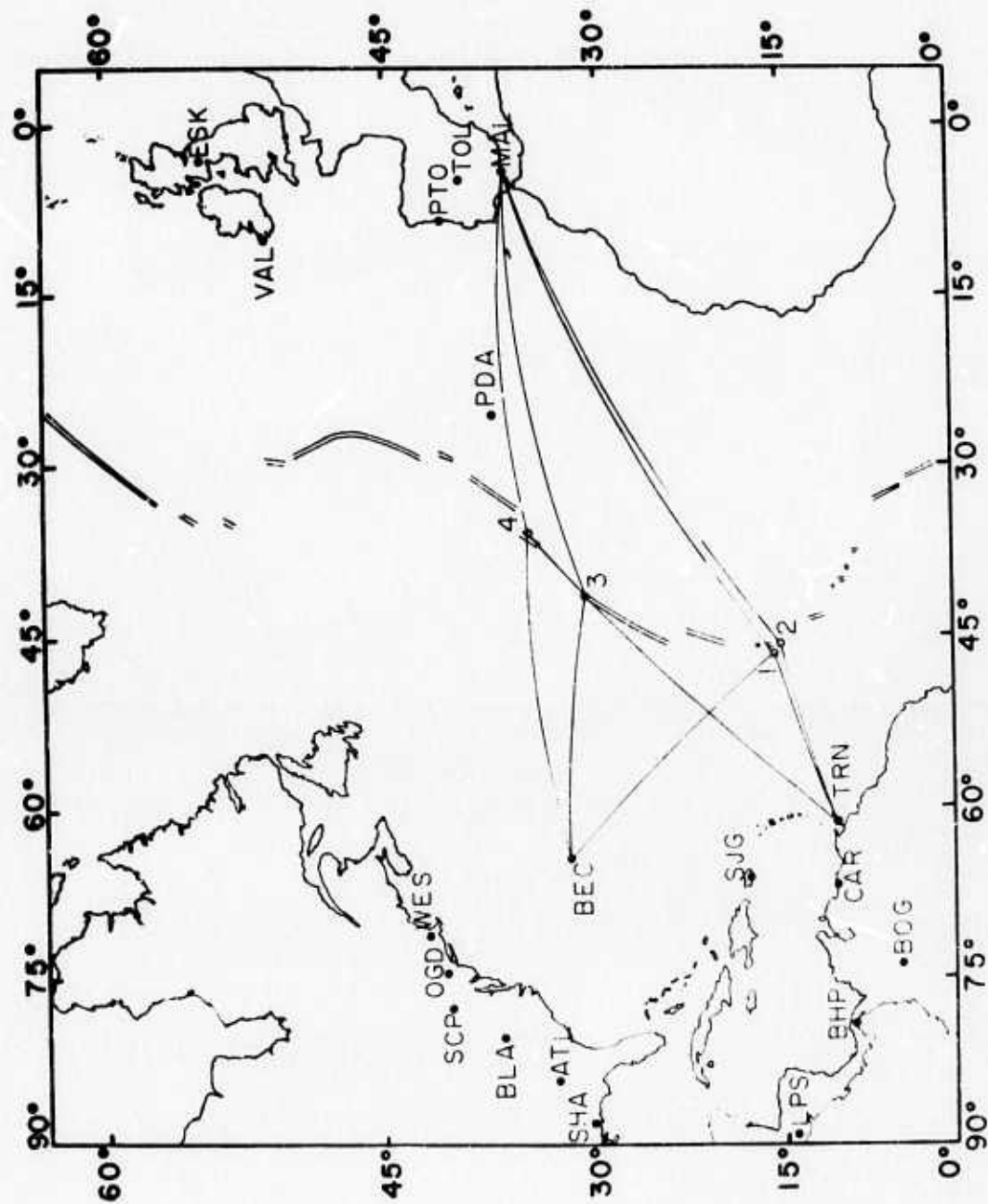


Figure 3.



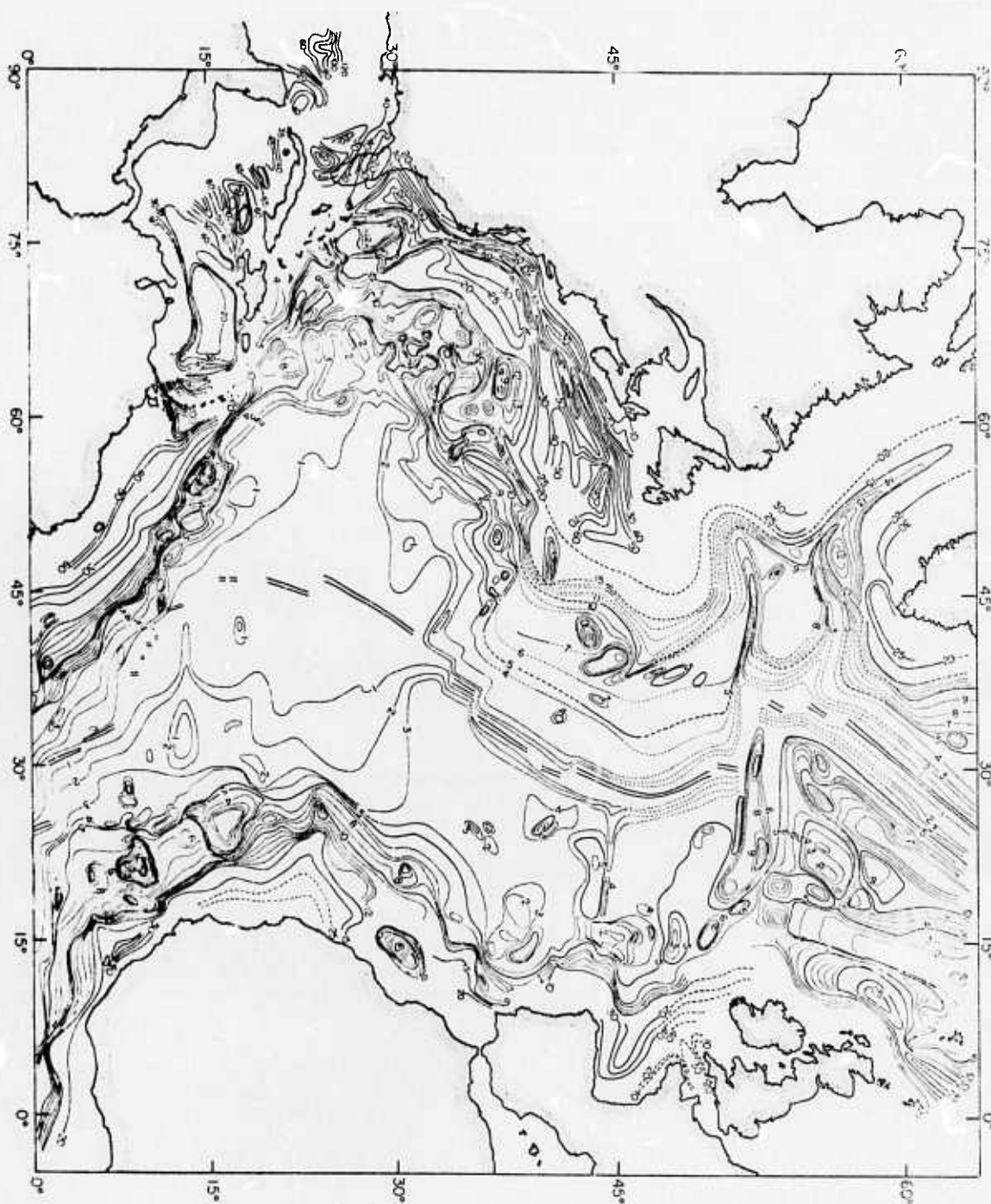


Figure 4.

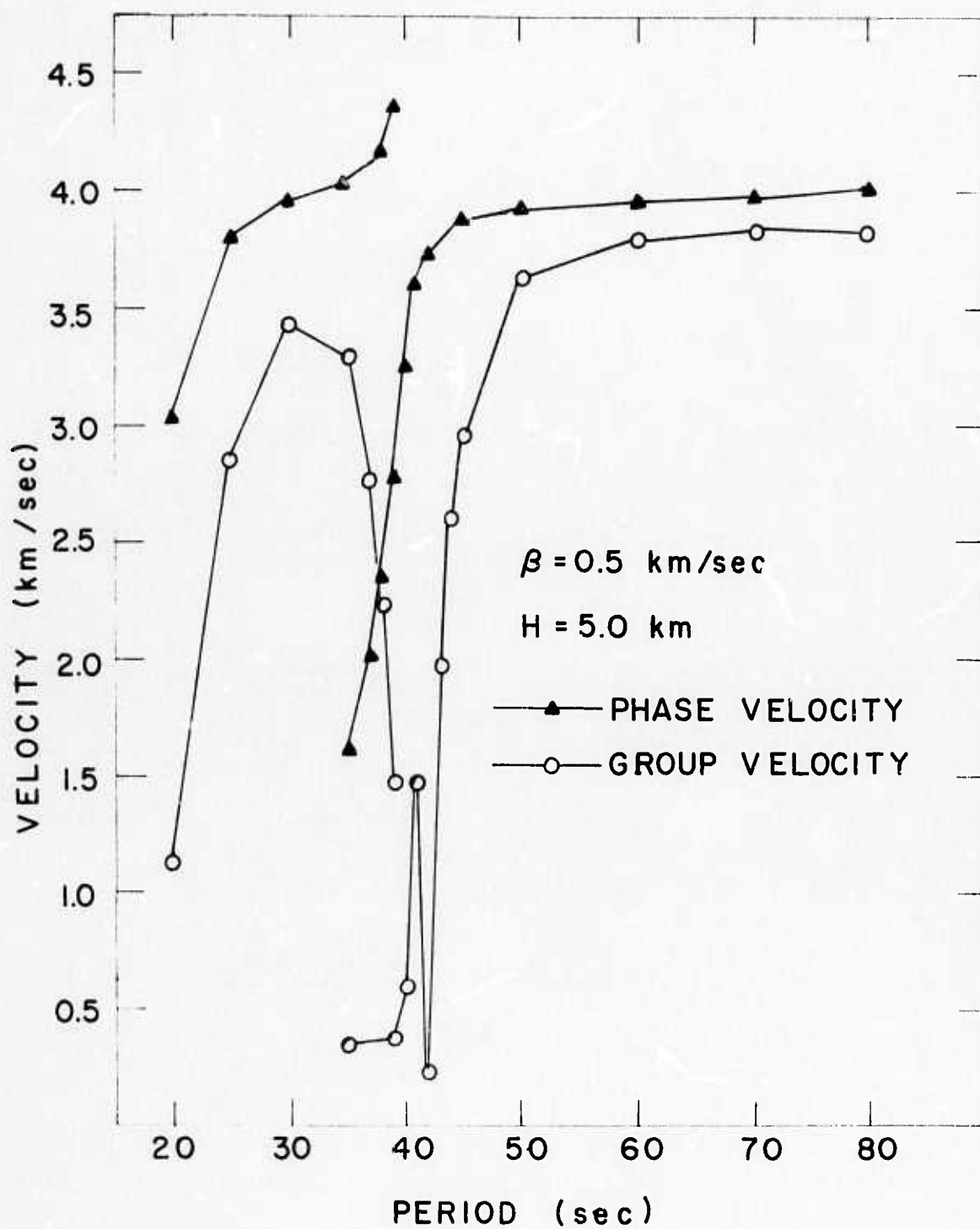


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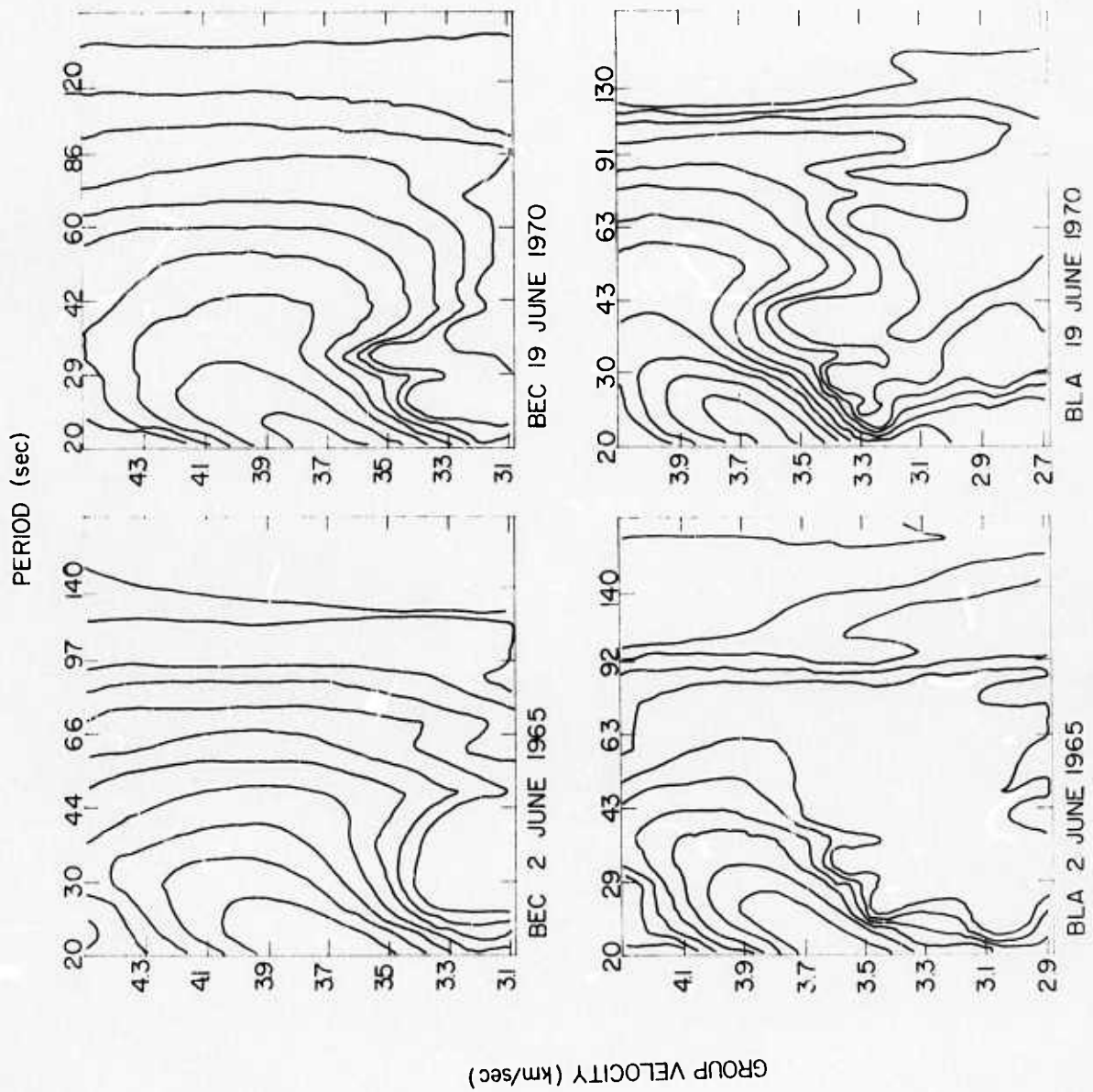


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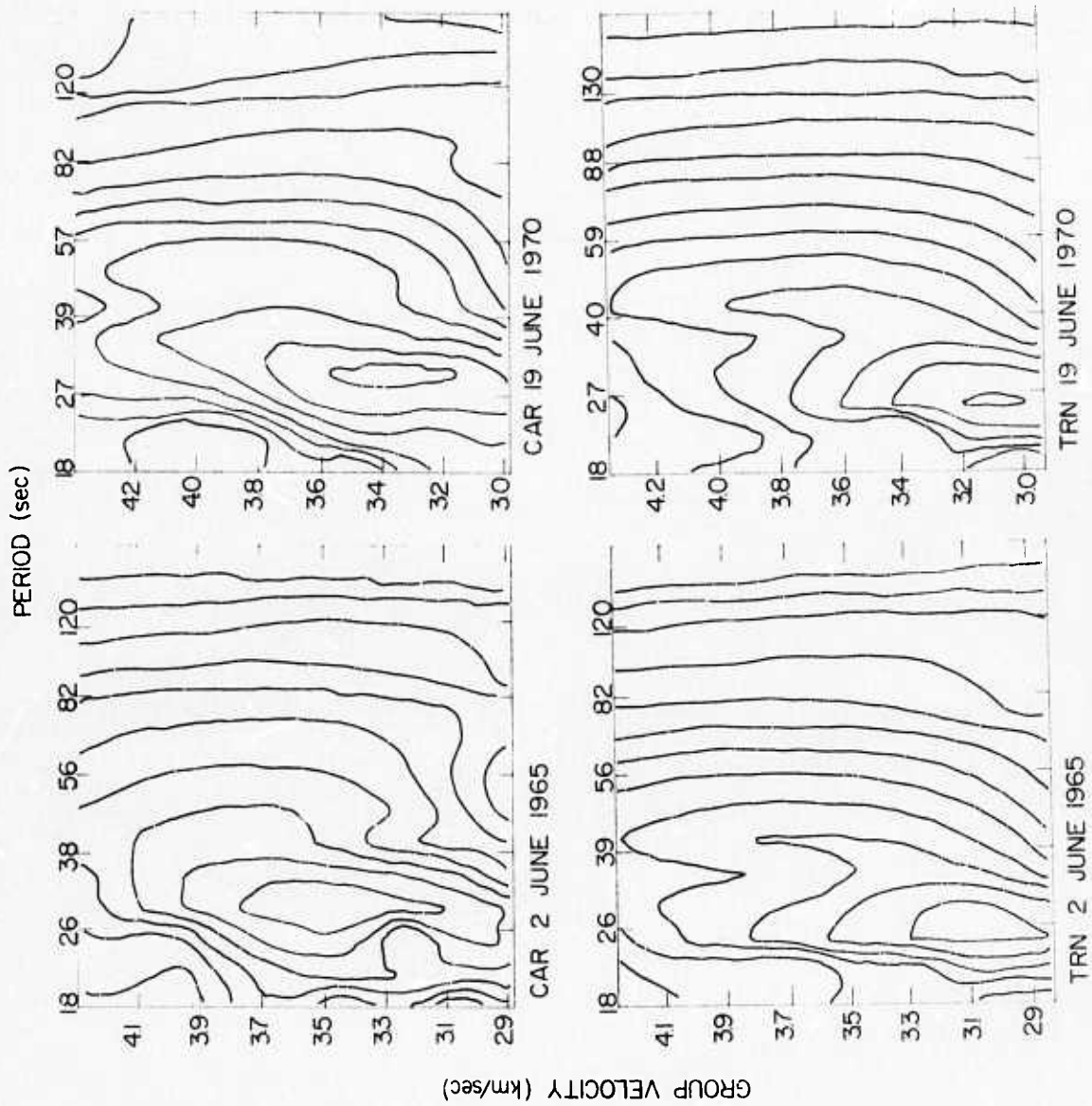
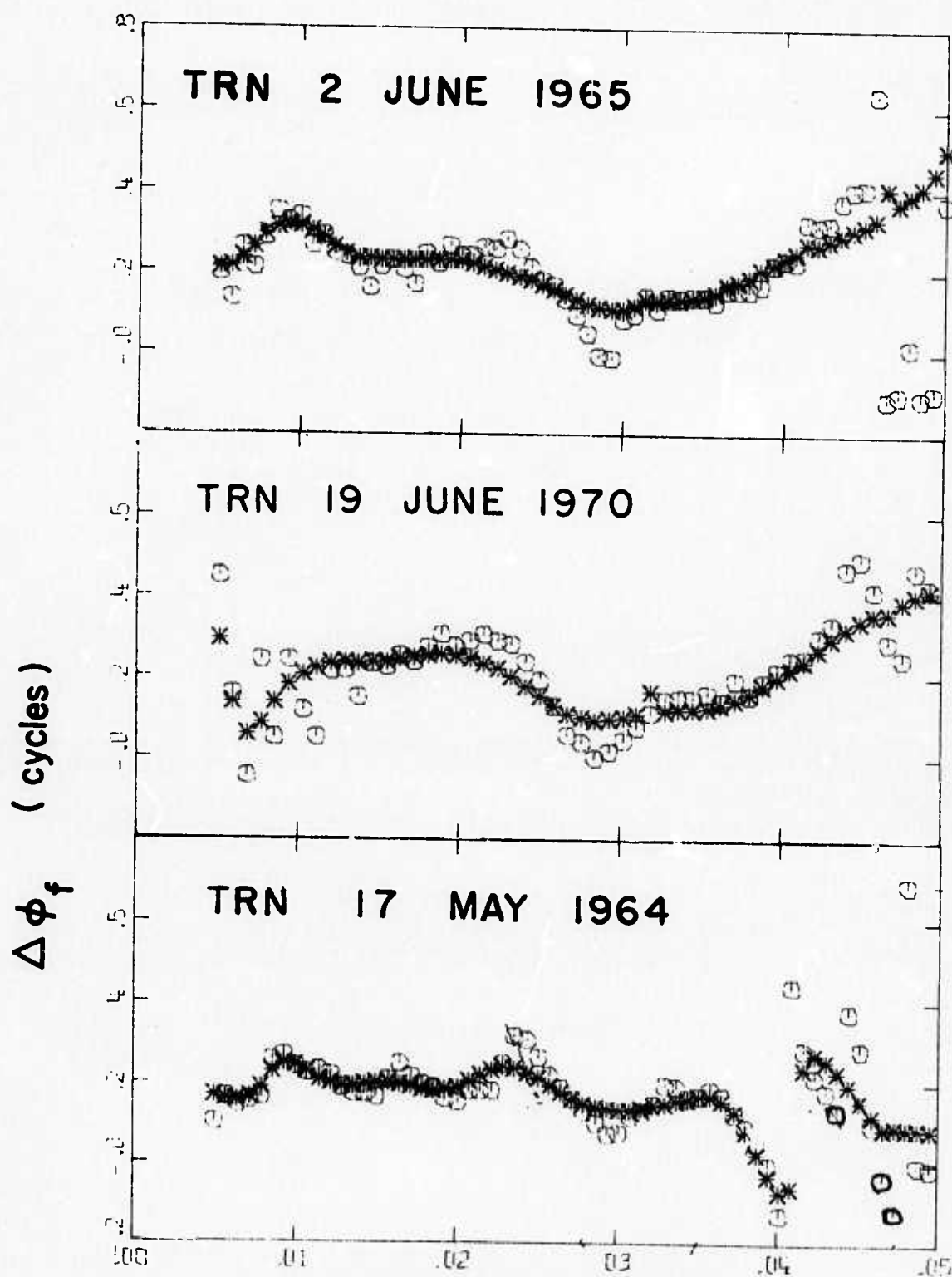
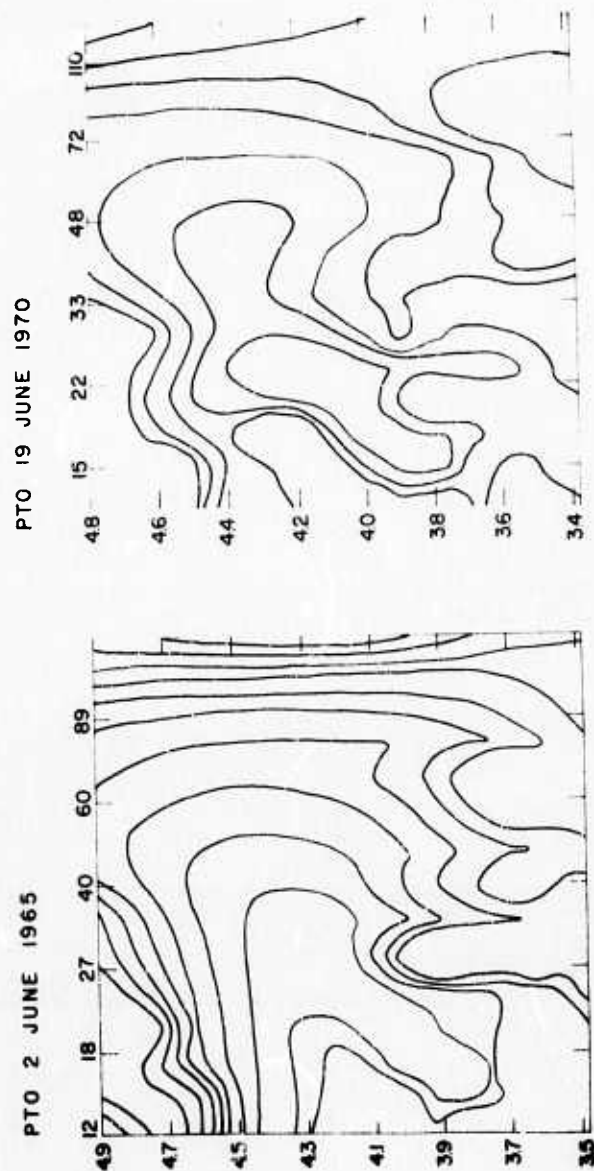
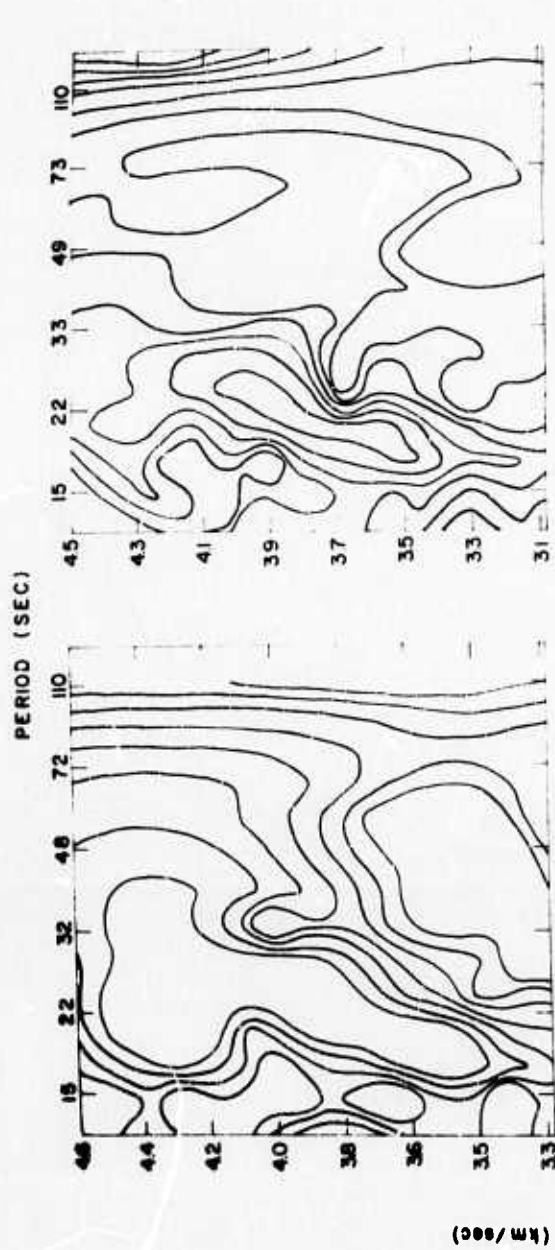


Figure 7.



FREQUENCY (Hz) Figure 8.



ATL 19 JUNE 1970

ATL 2 JUNE 1965

Figure 9.



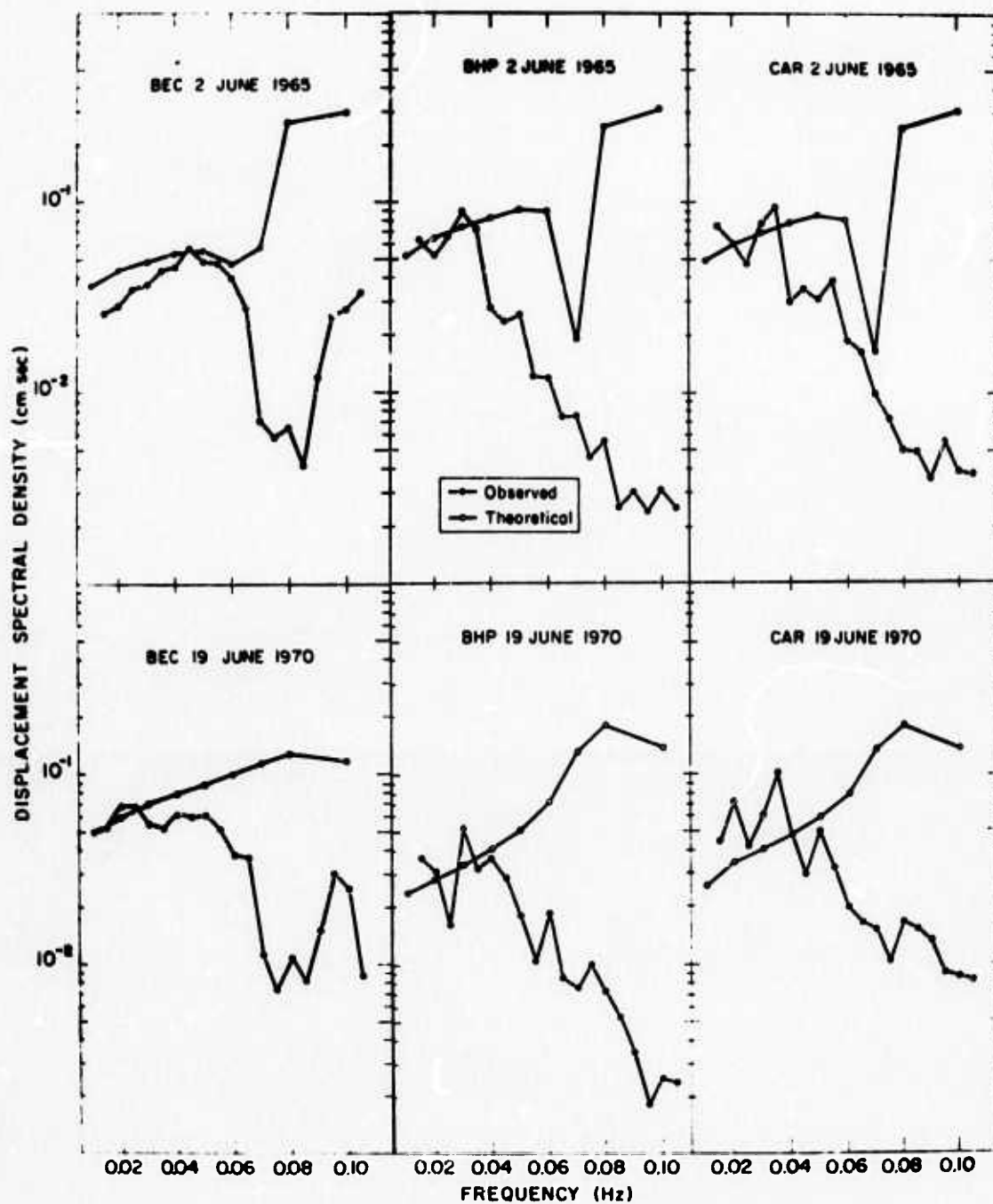


Figure 10.

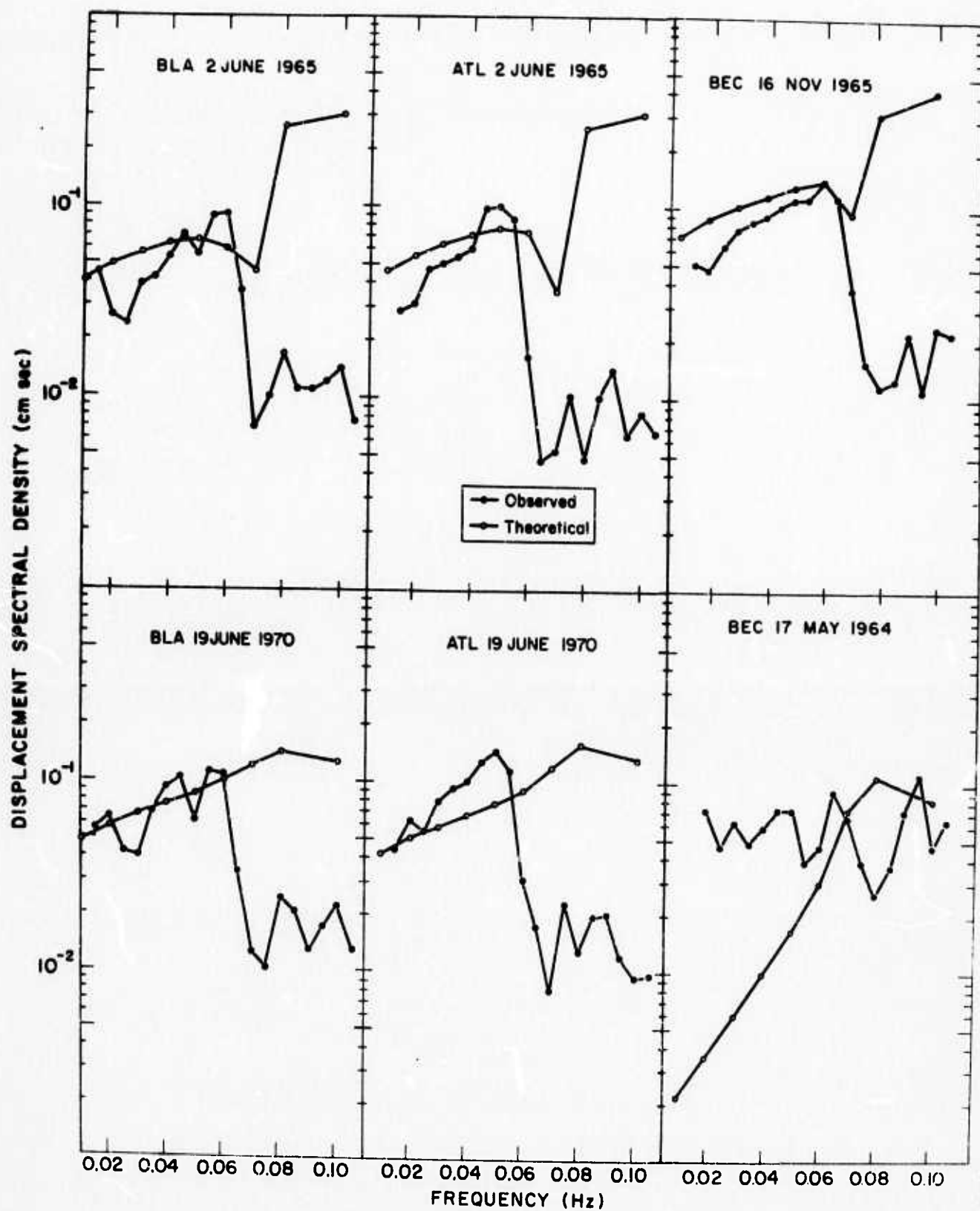


Figure 11.

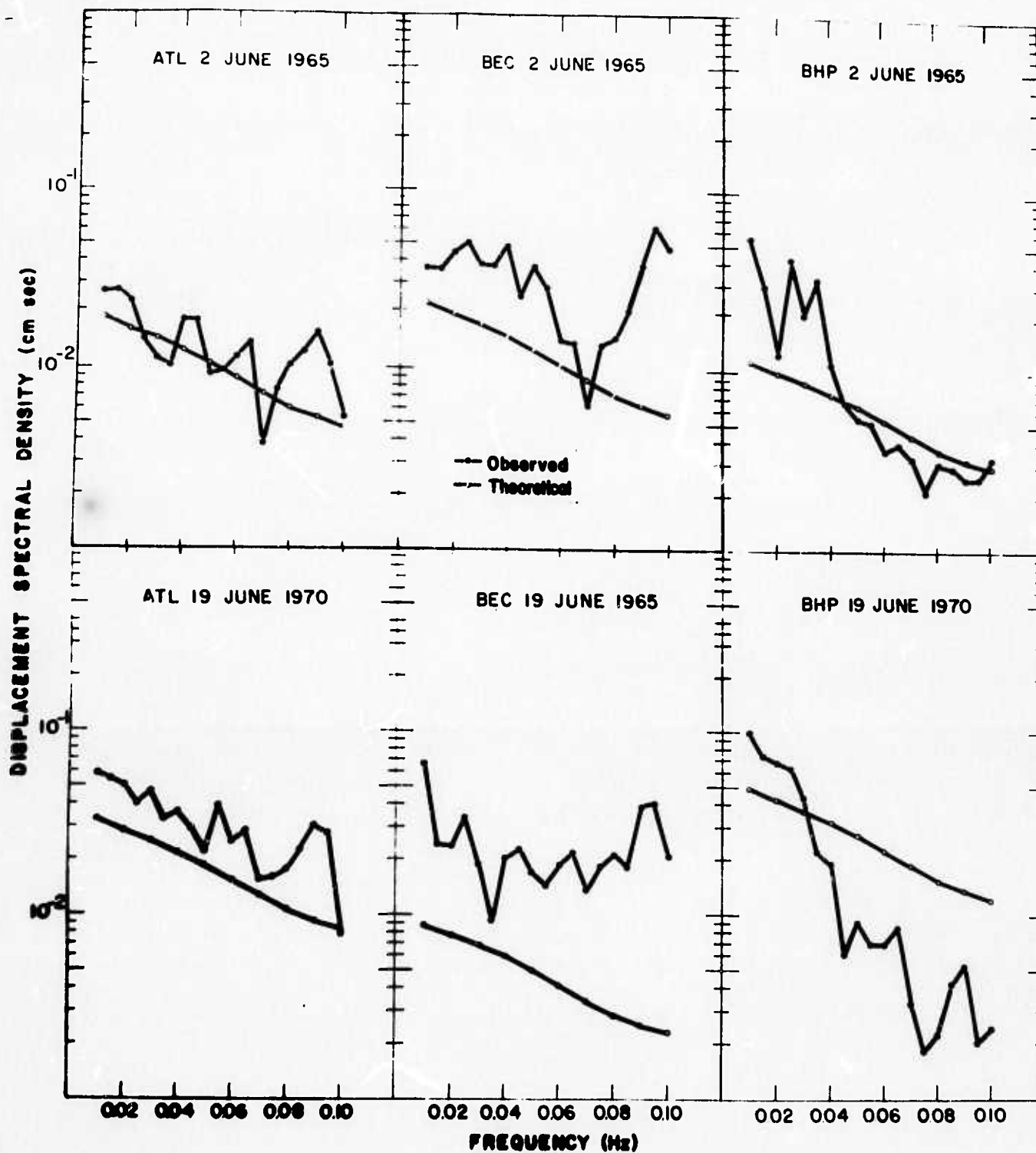
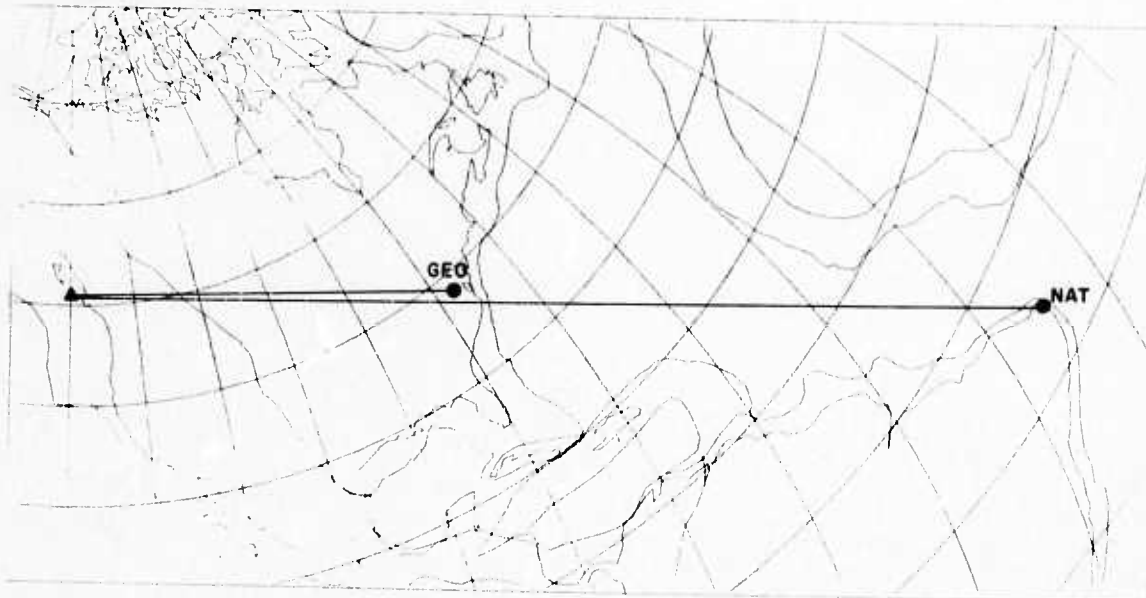


Figure 12.

29 APRIL 1967



4 FEBRUARY 1970

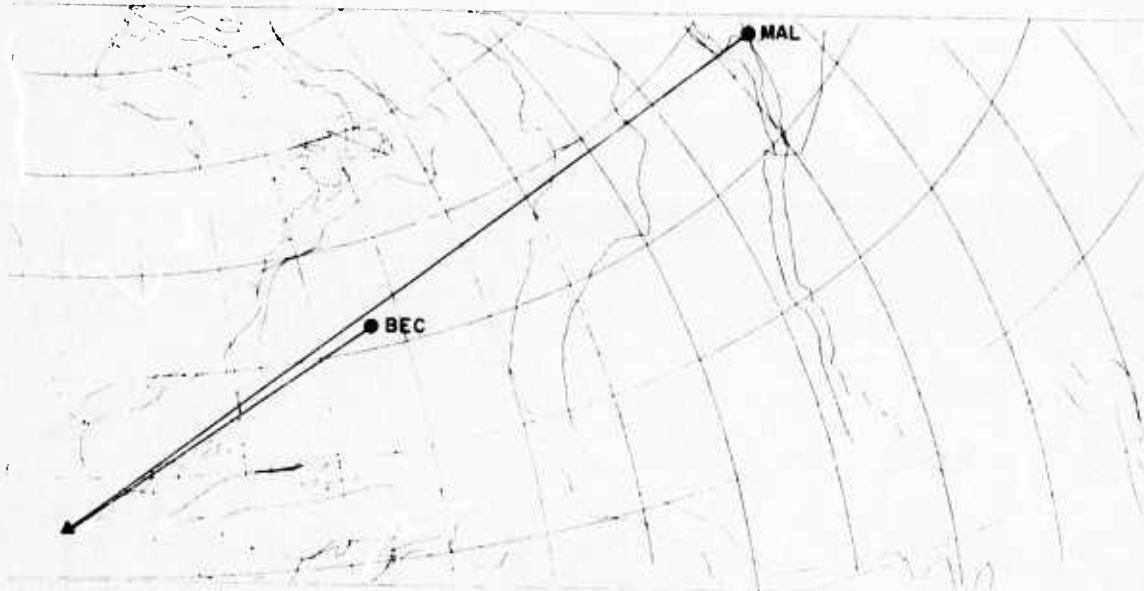
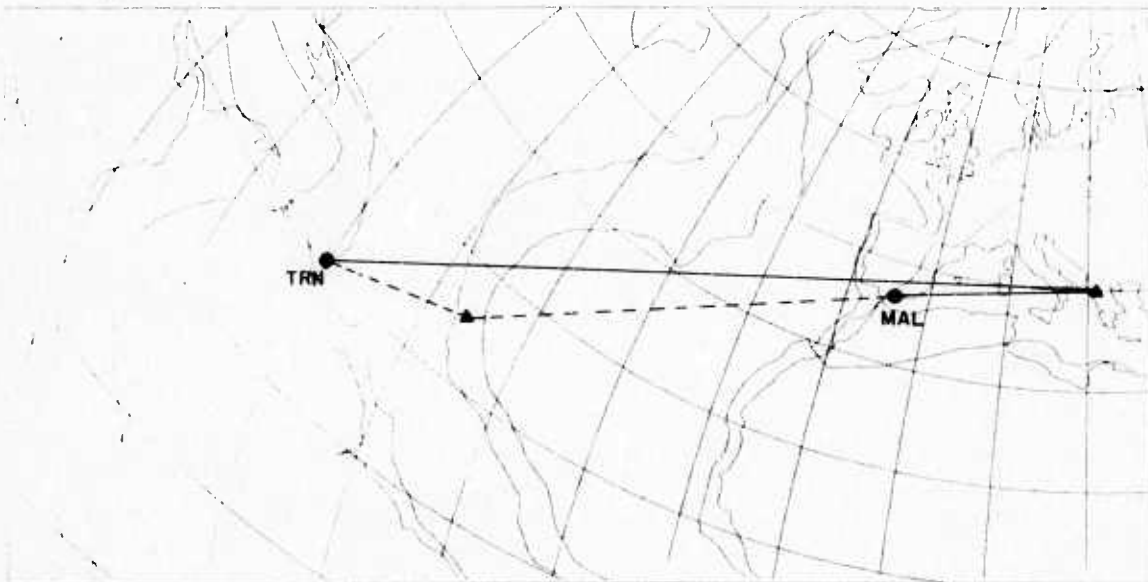


Figure 13.

1 MAY 1967



3 OCTOBER 1967

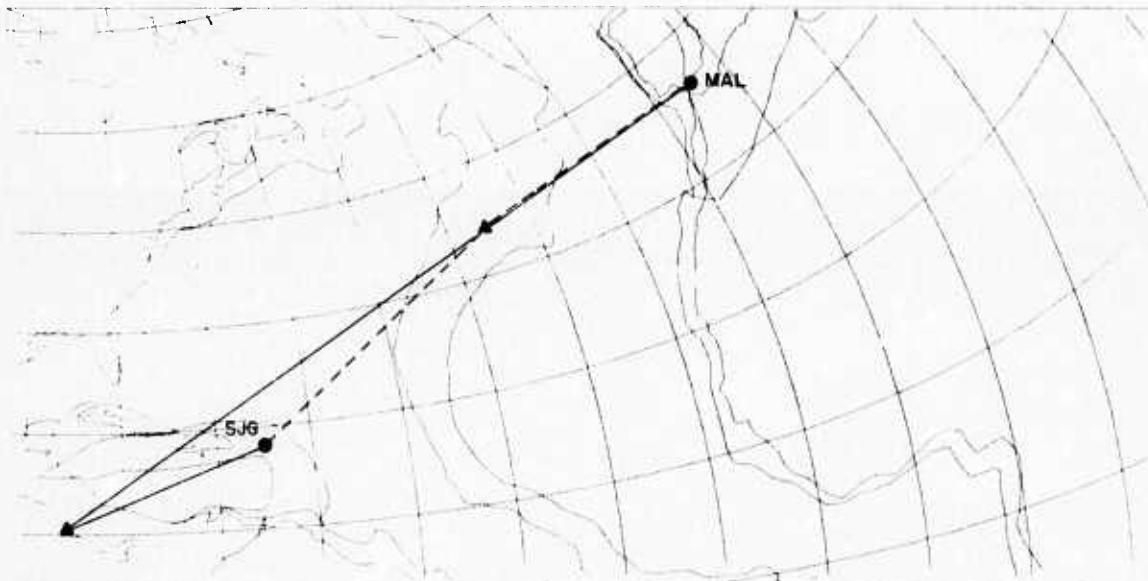


Figure 14.

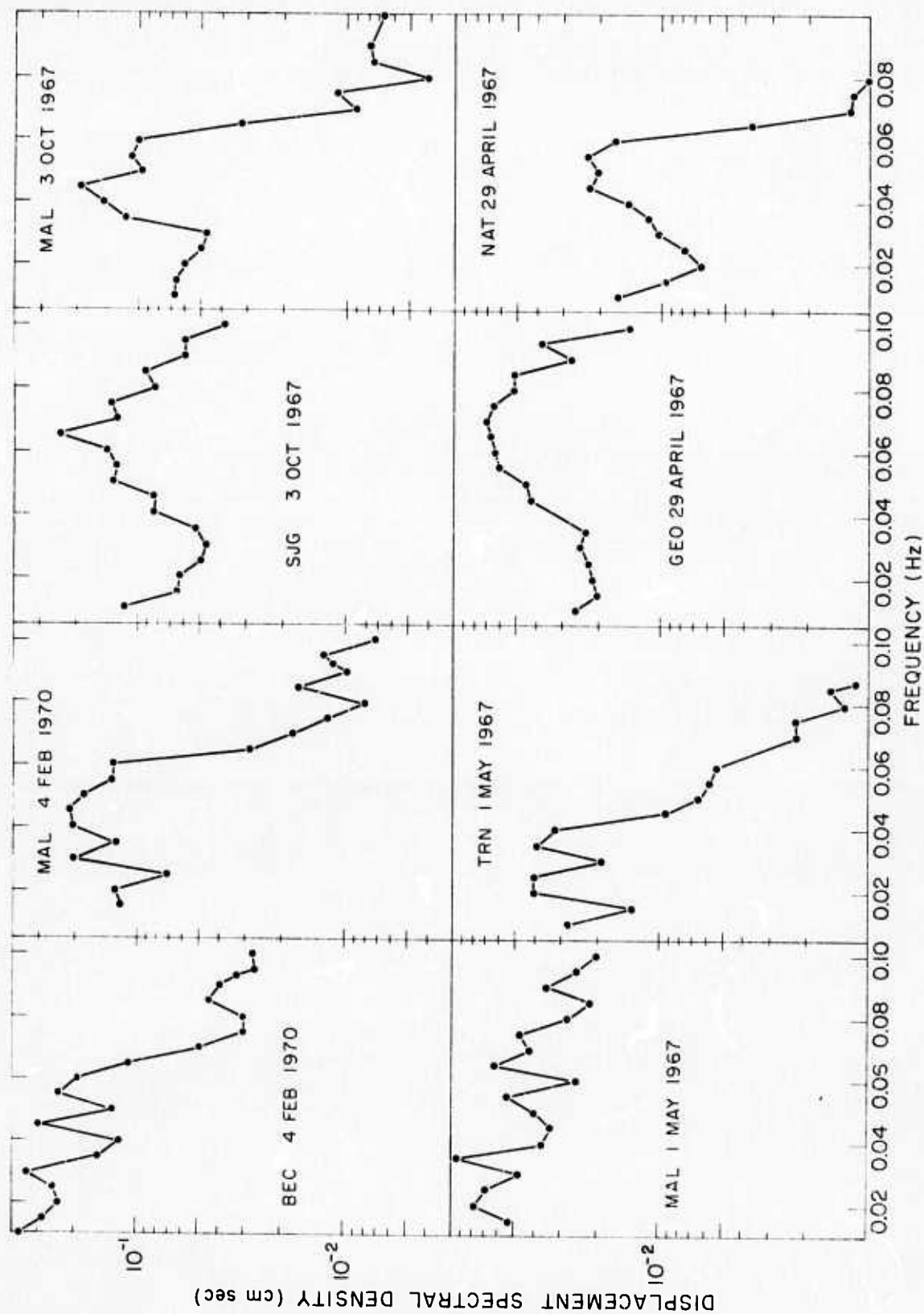


Figure 15.



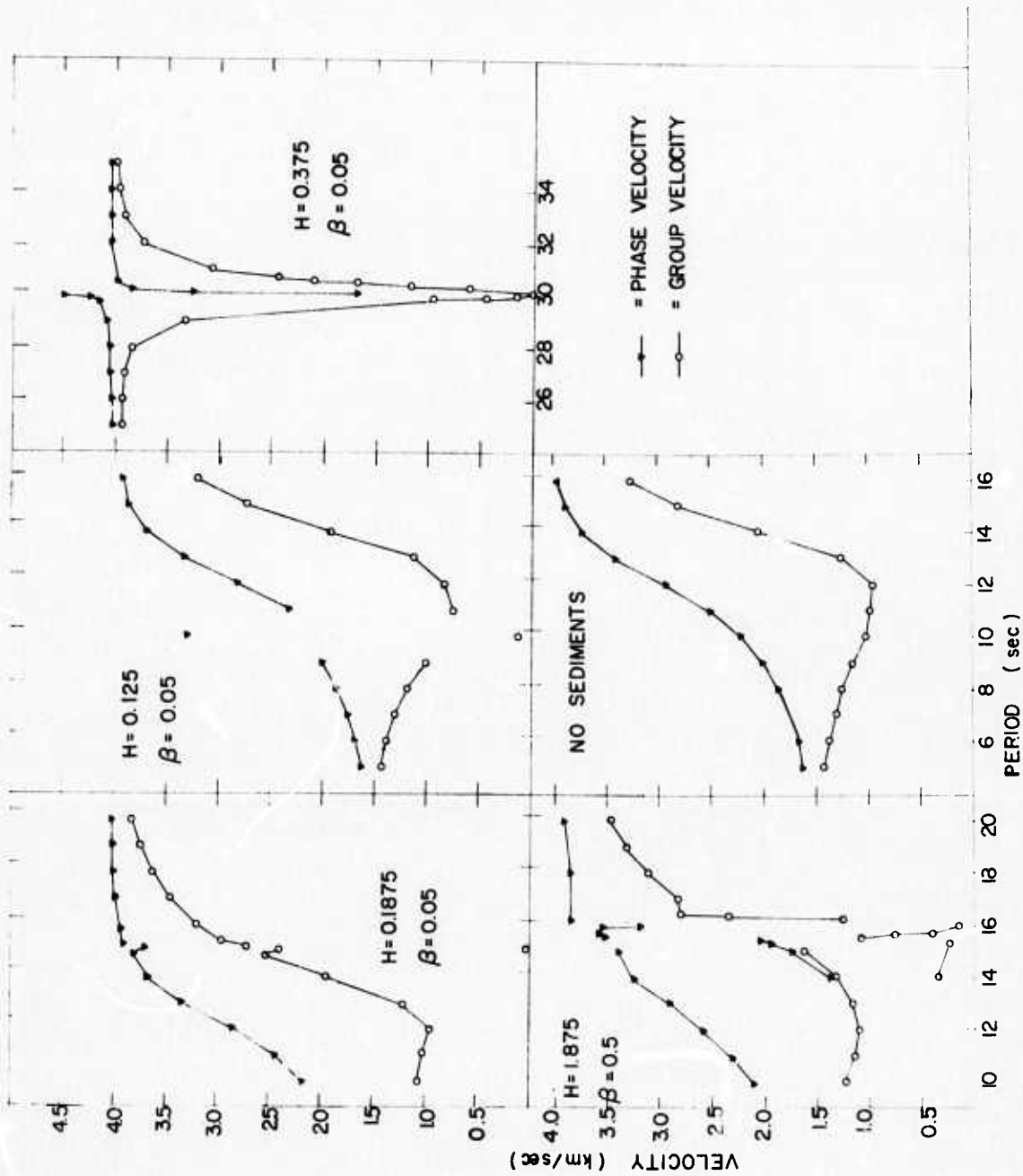


Figure 16.

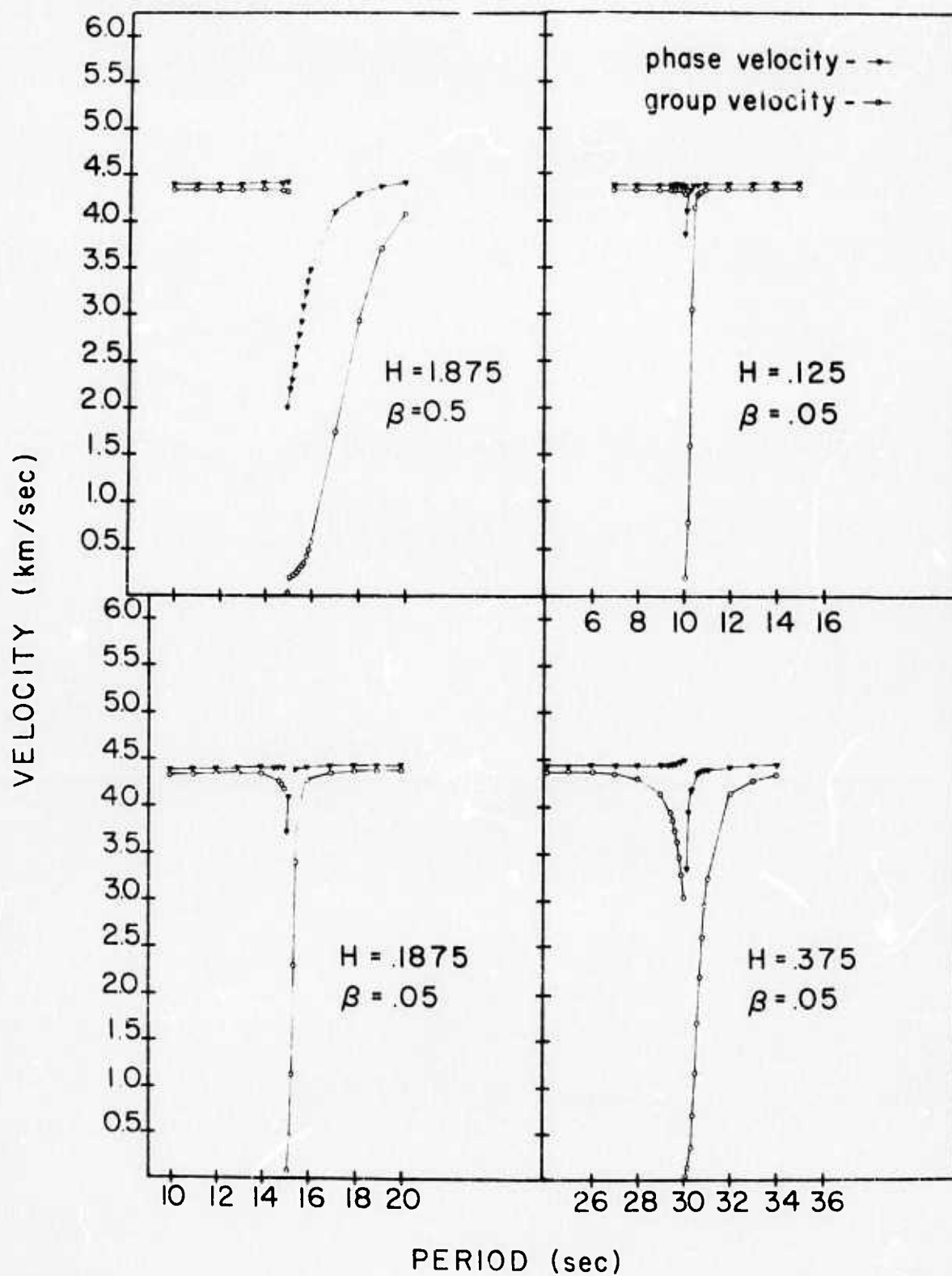


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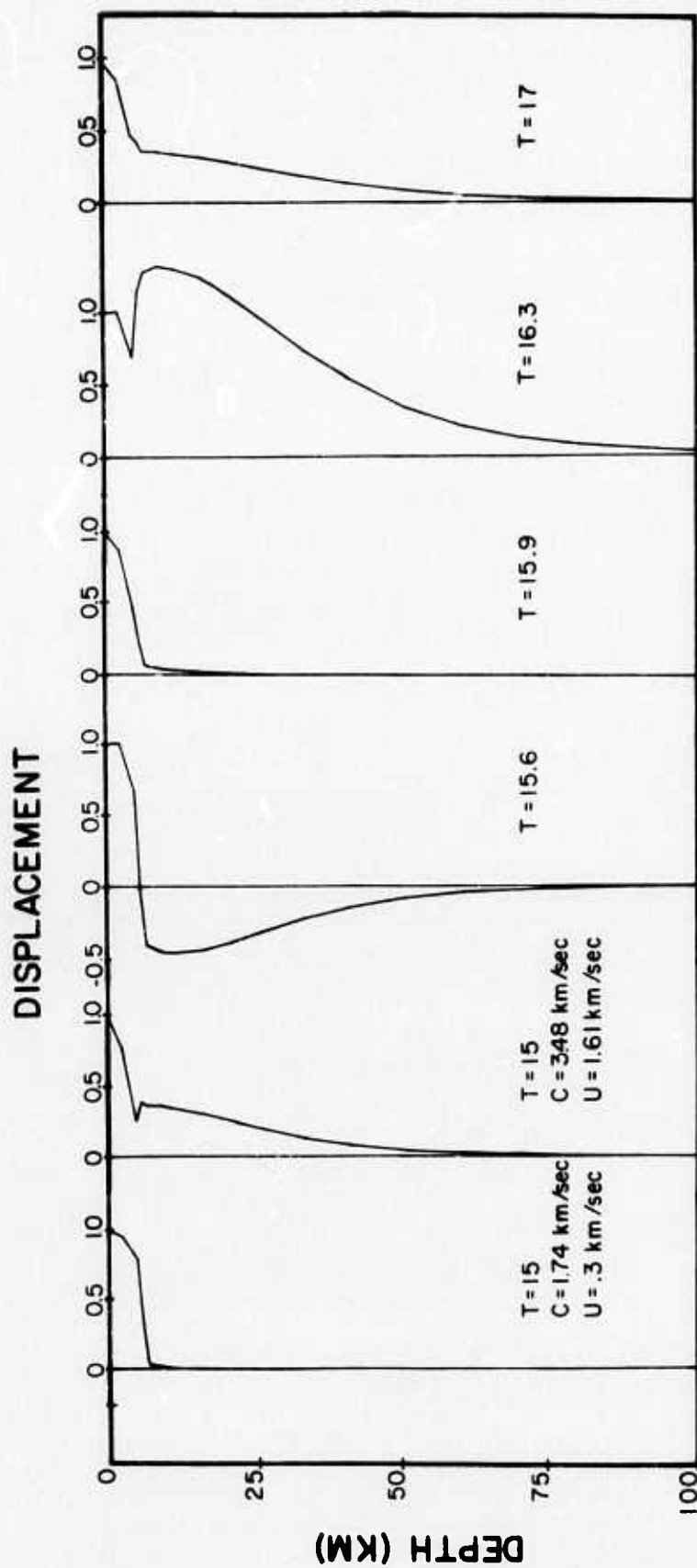


Figure 18.

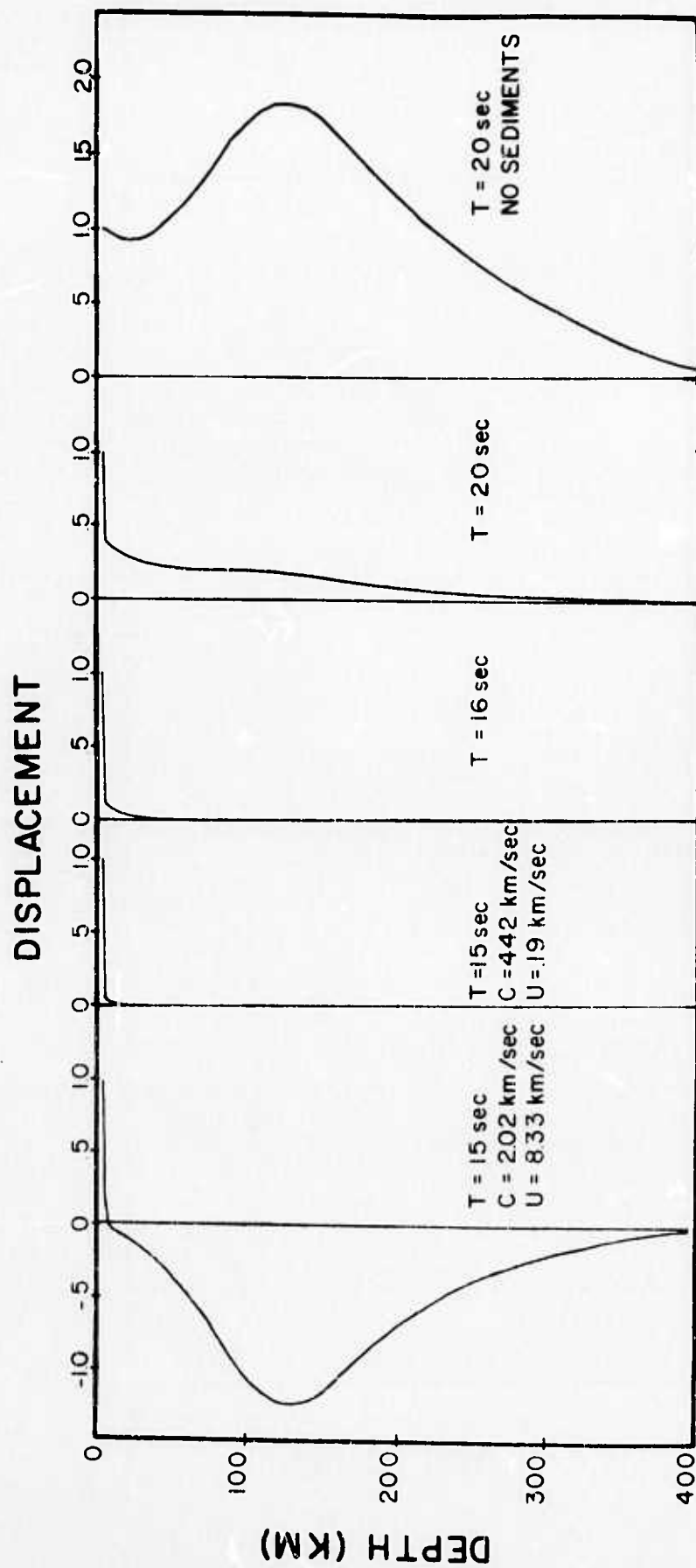


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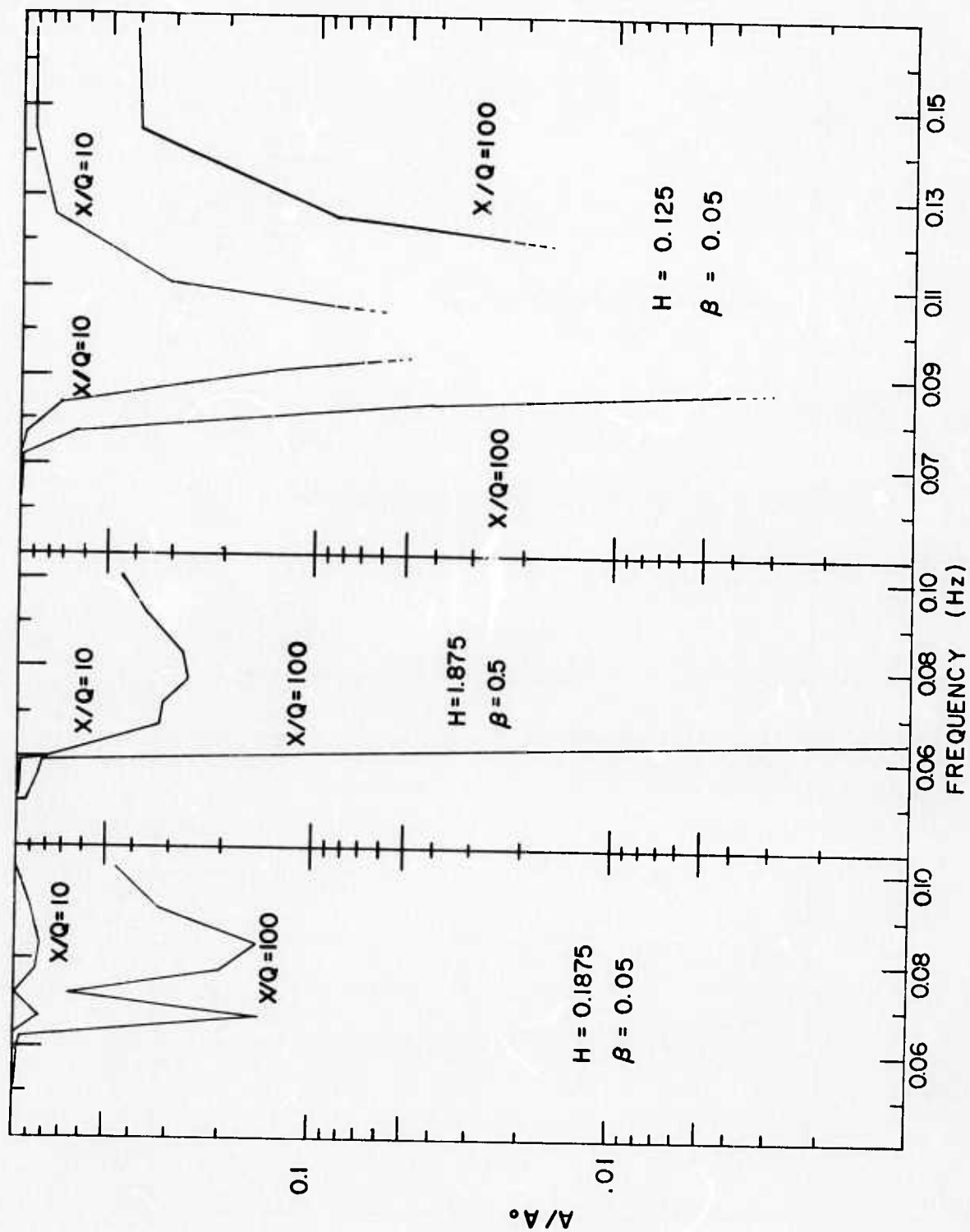


Figure 20.

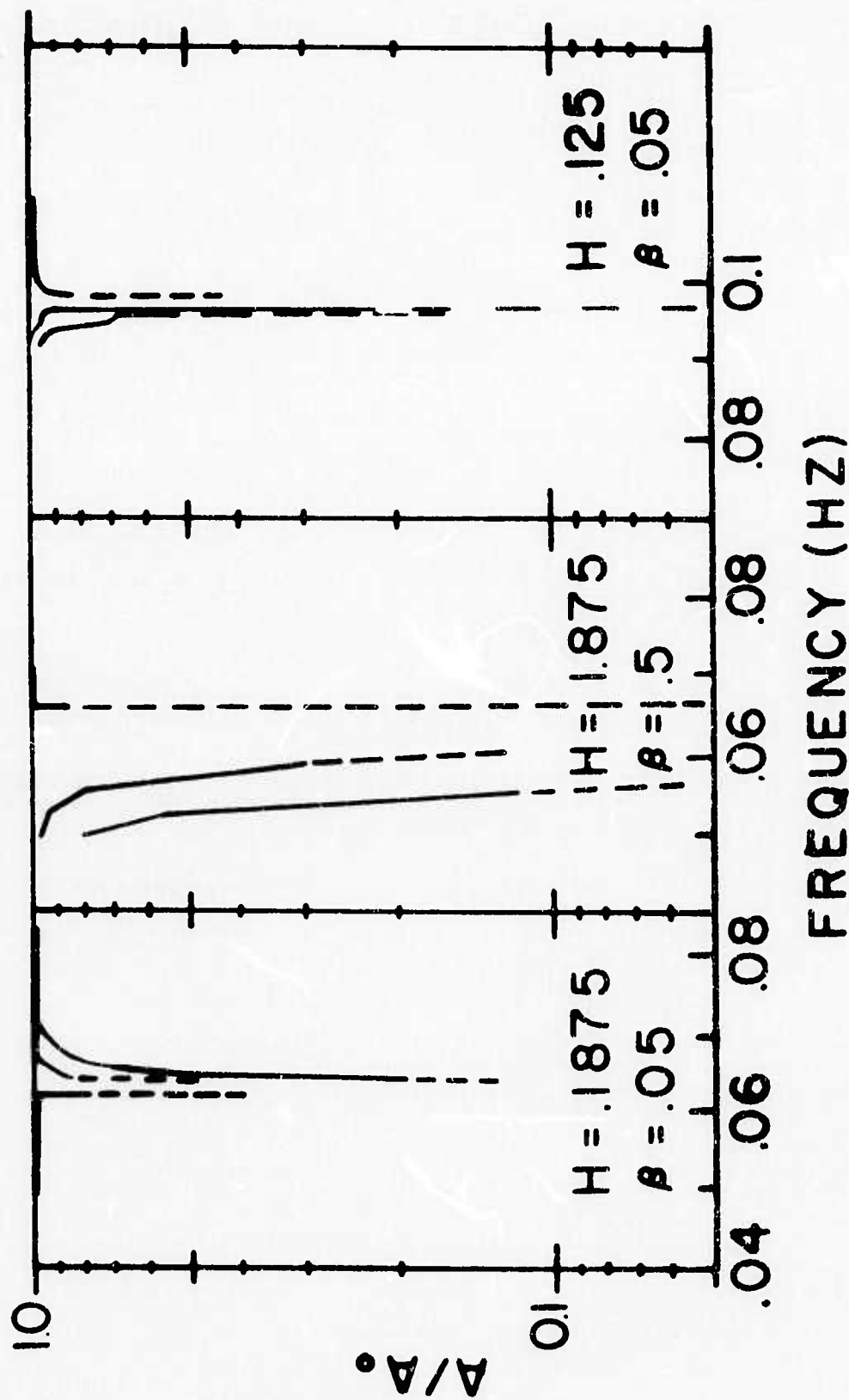


Figure 21.



PERIOD (sec)

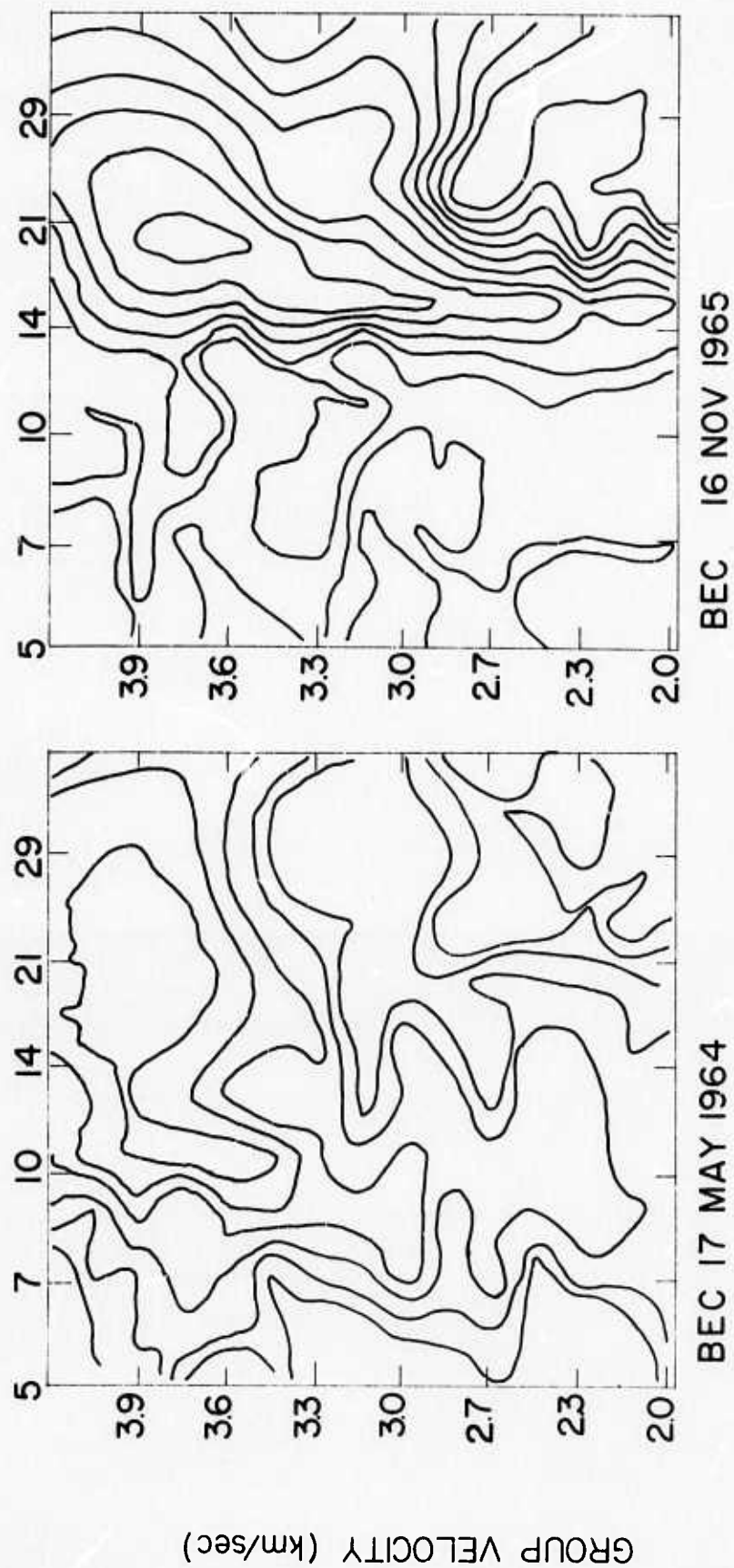


Figure 22.